Modeling the Seasonal Sea Ice Cycle in the Ross Sea, Antarctica

Yusuf Sinan Hûsrevoğlu
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MODELING THE SEASONAL SEA ICE CYCLE IN THE ROSS SEA, ANTARCTICA

by

Yusuf Sinan Hüsrevoğlu
Doctor of Philosophy

A Dissertation Submitted to the Faculty of Old Dominion University in Partial Fulfillment of the Requirement for the Degree of DOCTOR OF PHILOSOPHY OCEANOGRAPHY

OLD DOMINION UNIVERSITY
May 2008

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ABSTRACT

MODELING THE SEASONAL SEA ICE CYCLE IN THE ROSS SEA, ANTARCTICA

Yusuf Sinan Hüseveroğlu
Old Dominion University, 2008
Director: Dr. John M. Klinck

A mesoscale resolution (5 km) regional ocean model (ROMS), coupled thermodynamically/dynamically to a sea ice model (CICE) and thermodynamically to an ice shelf is used to investigate Ross Sea seasonal sea ice cycle, polynya dynamics, distribution and transformations of continental shelf water masses, and bottom water formation. Daily atmospheric forcing is from the ECMWF ERA-40 dataset, and for a separate simulation, wind forcing for Terra Nova Bay is substituted from daily automatic weather station (AWS) data. Simulated Ross Sea sea ice concentration reproduces the winter lead opening events observed in the SSM/I signal; however, it underestimates open water area (rmsd ~20%). Timing and advance of the Ross Sea spring-summer polynya are well captured. No local melting takes place during winter and over two years of simulation, heat loss at the ocean surface is offset 90% by lateral oceanic heat flux. Terra Nova Bay cumulatively produces more than twice the sea ice when forced with AWS winds. Forcing the Terra Nova Bay polynya with weaker winds result in continuous erosion of the High Salinity Shelf Water (HSSW) layer over the western Ross Sea continental shelf. Enhanced sea ice production and export driven by realistic winds are required to maintain the northward transport of dense shelf water.

High Salinity Shelf Water (HSSW) is formed in and exported from Terra Nova Bay and Ross Sea polynya areas at 0.14 and 0.64 Sv over two years of simulation. The larger area including the coastal polynya regions in the western Ross Sea provides a 1 Sv HSSW source while the Ross Ice Shelf (RIS) is a sink for about 0.4 Sv. Low Salinity Shelf Water (LSSW) outflow from beneath RIS cavity is 0.60 Sv. Modified Shelf Water (MSW)/Antarctic Bottom Water (AABW) is abundant over the entire continental shelf, forming the anticyclonic cell over the western Ross Sea. MSW/AABW net off-shelf transport and Modified Circumpolar Deep Water (MCDW)/Lower Circumpolar Deep Water (LCDW) net onshelf transport are 2.23 and 0.7 Sv, respectively.
Replacing AWS winds with ECMWF ERA-40 winds over the Terra Nova Bay results in larger scale dilution of HSSW in the depressions of the western Ross Sea shelf, diminishes HSSW circulation and transport northward along Victoria Land Coast, disrupts the western gyre, and causes an overall decrease in vertically averaged transport over the western Ross Sea shelf.
To my family
ACKNOWLEDGMENTS

I would like to thank my advisor Dr. John M. Klinck for his guidance, patience, and mentorship. I am very grateful to Dr. Eileen E. Hofmann, who, together with Dr. Klinck, provided continuous support and encouragement to make this study possible. I extend many thanks to the committee members, Drs. Muench, Grosch, and Adam, for their reviews, comments, and suggestions on the manuscript.

Many members of the Center for Coastal Physical Oceanography (CCPO) at Old Dominion University contributed to the completion of this study. Mike Dinniman, always the most helpful, deserves special recognition for providing the foundation to the ocean model application used in this study. Many many thanks to Julie Morgan, who worked diligently to handle all administrative and financial matters related to this project. Gabriel Franke, friendly and cheerful, made a great difference in making CCPO resources easily accessible to all graduate students. Joe Ruettgers, there for computers and beyond, was helpful when one knew how to ask.

And my fellow CCPO graduate students, friends: Erik Chapman, a true comrade, has always been there to support and help, especially when most needed. I am grateful to Erik for his companionship, ideas, and conversation which, along with his family, Michele, Reece, and Kira, are already greatly missed. Diego and Andrea, warm, mellow, and with sense of humor, showed up later in time, but our friendship took off fast. I cannot thank them enough for their support in the final stages and, Diego, for his help months after my departure.

Many other dear friends showed great companionship and support during this period. The Duran family, Fato§, Hakan, Sila, Neriman, and Cemal became a real family to me, feeding taking care of me Turkish style or otherwise. Pınar Özduration was there in almost all crisis situations that involved an interview, travel, or moving. İlker Türkgedi, selfless and pure, always offered help without asking and provided it when most needed. Elif Demir continued to be a close friend from a distance always conveying encouragement. Bayram Çelik and Didem Özden were sources of intellectual and emotional insight. Elizabeth Aucamp of cheer, and Mert Çadircu, my final roommate, of help during and beyond my departure from Norfolk. The named few and the majority not named here made my overall experience in the U.S. a great one.

I also would like to thank to all those who contribute to and make publicly available the community models, earth system datasets, programming utilities, and open
source operating systems and software. This study was funded by the National Science Foundation Grant OPP-03-37247. The computer facilities and resources were provided by the Commonwealth Center for Coastal Physical Oceanography at Old Dominion University.
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CHAPTER I
INTRODUCTION

I.1 ROSS SEA SEASONAL SEA ICE CYCLE

Seasonal sea ice cycle and dense water formation are essential and distinguishing characteristics of high latitude oceans. Shelf and surface waters are ventilated by atmospheric cooling and densified by brine rejection from sea ice formation [Jacobs, 2004]. Gaining negative buoyancy as a result of these interactions, waters in the upper layer sink, mixing with subsurface and deep waters, and subsequently drive the global thermohaline circulation [Orsi et al., 2002]. Seasonal sea ice growth and distribution by ice motion and deformation therefore influence dense water production [Kwok, 2005]. Varying seasonally in extent and covering about 7% of ocean surface at any given time [Parkinson and Washington, 1979], sea ice is also an important component of the global climate system [DeLiberty et al., 2004]. Seasonal sea ice cover forms an insulating boundary and increases surface albedo, and as such, reduces the atmosphere-ocean heat, mass, and momentum exchange [Assmann, 2003].

Antarctic sea ice is thinner than its Arctic counterpart and covers 20% more area at its maximum extent [DeLiberty et al., 2004]. Seasonal variation in sea ice area is about 80%, which also is larger than that for the Arctic [Schellenberg et al., 2002]. Studies suggest that interannual variability in the Antarctic sea ice cover may be connected to sensitivity to atmospheric and oceanic forcing [Geiger et al., 1997] as well as global and circumpolar phenomena such as the Antarctic Circumpolar Wave (ACW) [White and Peterson, 1996], El Niño-Southern Oscillation (ENSO) [Yuan and Martinson, 2000], and the Southern Annular Mode (SAM) [Hall and Visbeck, 2002]. Dense water produced in the Southern Ocean as a result of atmosphere-sea ice-ocean interaction is the primary source of bottom waters in the abyssal ocean [Jacobs, 2004]. Therefore, investigation of the coupled sea ice-ocean system in the Antarctic is essential to understanding its sensitivity to and feedback on the global climate.

In the context of Antarctic studies, the Ross Sea (Figure 1) has been recognized since late 1970s as an important source region for Antarctic Bottom Water (AABW) [Jacobs, 2004]. Featuring a large ice shelf cavity (surface area ~500,000 km²), a wide continental shelf, recurring polynyas, and a narrow and steep continental slope, the

This dissertation follows the style of Journal of Geophysical Research.
Figure 1. Map of the Ross Sea. The inset (after Kurtz and Bromwich [1983]) shows the geographical details of the Terra Nova Bay with the bold arrow indicating the median wind direction. Indicated on the map of the Ross Sea region (after Kurtz and Bromwich [1985]) are the locations of automatic weather stations (AWS), a National Centers for Environmental Prediction (NCEP) grid node, and an International Satellite Cloud Climatology Project (ISCCP) grid node. Figure reproduced from Van Woert [1999b].

region has been the subject of an increasing number of observational and modeling studies. This study aims to investigate the coupled sea ice-ice shelf-ocean dynamics of the Ross Sea using regional numerical models with spatial resolution high enough to resolve coastal polynyas and small-scale bathymetric features.
The Ross Sea is on the average more than 80% ice covered for up to 240 days of the year [Jacobs and Comiso, 1989]. Most of the ice cover consists of annual pack ice with total area extent ranging from $4 \times 10^6$ km$^2$ in summer to about $20 \times 10^6$ km$^2$ in late winter, i.e. during September-October [Cavalieri et al., 1999]. Sea ice increases surface albedo, limits air-sea gas exchange, and furnishes a habitat for microbial communities [Arrigo et al., 2003]. Sea ice cover is strongly seasonal and its thickness is limited to between 0.5 m and 1.0 m because of melting due to the entrainment of warm and salty Circumpolar Deep Water (CDW) into the mixed layer [Gordon and Huber, 1990]. Jacobs and Comiso [1989] outline the primary characteristics of the large scale sea ice distribution in the Ross Sea. According to their study, the continental shelf is ice-free in late summer except for the ice field on the northeastern shelf. Sea ice grows around this residual field and along the coastline in autumn and reaches high concentrations, although the northwestern shelf remains ice-free late into March. Winter sea ice concentrations over the continental shelf and along the coastline are lower compared to that of the oceanic region. By early December a large and several smaller ice-free areas within the pack ice, termed polynyas, are formed on the western Ross Sea shelf (Figure 1).

Ocean circulation and sea ice dynamics in the Ross Sea have been investigated extensively through observational studies (e.g., Jacobs et al. [1985]; Jacobs and Comiso [1989]; Locarnini [1994]; Arrigo et al. [1998]; Jacobs and Giulivi [1998]) and programs (e.g., Climatic Long-term Interactions for the Mass-balance in Antarctica (CLIMA) of the Italian National Program for Antarctic Research (PNRA) [Budillon et al., 2000], the United States Joint Global Ocean Flux Study (US JGOFS) in the Southern Ocean, Antarctic Environment and Southern Ocean Process Study (AESOPS) [Smith et al., 2000]), and modeling studies. These studies provided insight into the sea ice dynamics and circulation in the Ross Sea basin. However, the coupled three-dimensional sea ice-ocean dynamics problem at the spatial resolution high enough to resolve smaller scale processes (e.g., time evolution of polynya area, frazil ice drift, realistic representation of bathymetry, water masses, and circulation) has not been carried out [Morales Maqueda et al., 2004]. The first part of this study will focus on the following research questions: What are the important processes and forcing that determine the seasonal, large-scale sea ice cover in the Ross Sea? What causes the regional differences in sea ice formation and export? How is the sea ice mass balanced?
I.2 POLYNYA DYNAMICS

A polynya is defined as "any non-linear shaped opening enclosed in ice" which "may contain brash ice and/or be covered with new ice, nilas or young ice" [World Meteorological Organization, 1970]. Polynyas can occur in winter when air temperatures are sufficiently low for sea water to freeze [Smith et al., 1990; Van Woert, 1999a]. They are predictable and recurrent phenomena occurring in the same region, typically rectangular or elliptical, bordered by ice covered waters or land, and range in width from a few hundred meters to hundreds of kilometers [Smith et al., 1990]. Polynya dynamics play an important role in ocean-atmosphere heat exchange, sea ice production, dense shelf water formation, disintegration of ice pack in spring, and primary and secondary production in the polar oceans [Bromwich et al., 1998]. Polynyas increase oceanic heat and moisture losses to the atmosphere, ice and brine formation rates, and biological productivity [Jacobs and Comiso, 1989; Arrigo and McClain, 1994]. Open water contributes to localized spring phytoplankton blooms, and increased biological activity is implied by the presence of large mammals using polynyas as feeding grounds [Smith et al., 1990]. Heat exchange at the open water surface is two orders of magnitude greater than that through snow covered sea ice; therefore, polynyas dominate regional heat budgets in winter [Smith et al., 1990].

Mechanisms for polynya formation and maintenance vary. Areas where newly formed ice is advected out of the region of formation by winds and currents are called "latent heat" polynyas [Smith et al., 1990; Fichefet and Goosse, 1999]. Heat required to balance loss to the atmosphere and hence maintain an ice-free area is provided by latent heat of fusion as sea ice continually forms at the surface [Smith et al., 1990]. Latent heat released due to ice formation keeps the water at the freezing point and, therefore, does not affect freezing rates [Morales Maqueda et al., 2004]. Analytical and numerical models of latent heat polynya dynamics suggest that the polynya size and the resulting open water fraction are driven by the balance between ice advection at the polynya boundary and ice production within the polynya [Pease, 1987; Ou, 1988; Darby et al., 1995; Van Woert, 1999a]. Latent heat polynyas, due to rapid growth and removal of sea ice, are sites of active brine rejection which initiates the convection of denser surface water [Jacobs et al., 1995]. This process along with associated cooling of the ocean surface contributes to formation of deep water in the Southern Ocean [Killworth, 1983; Zwally et al., 1985; Cavalieri and Martin, 1985; Tear et al., 2003]. Deep convection associated with brine rejection plays a significant role in stable stratification and the climate of the world ocean by allowing heat to be
given up from deep ocean to the atmosphere [Killworth, 1983]. As opposed to land, ice formation on the ocean surface leads to a negative feedback mechanism when resulting deep convection melts some of the ice and decreases surface albedo, allowing surface ocean to get warmer [Killworth, 1983].

“Sensible heat” polynyas, on the other hand, are formed by the supply of oceanic heat to the surface waters which prevents local ice formation or melts the sea ice that has already formed [Smith et al., 1990; Fichefet and Goosse, 1999]. Sensible heat supplied to the surface ocean in the polynya area controls the freezing rate and the size of the polynya [Morales Maqueda et al., 2004]. However, Bailey and Lynch [2000] argue that the relative importance of ice melt due to deep ocean convection and ice advection out of the area by atmospheric or oceanic forcing in the initiation and maintenance of sensible heat polynyas has not yet been fully understood. Deep water polynyas are generally driven by sensible heat and occur where warmer, saltier deep water is separated by a weak pycnocline from the cold, fresh surface layer [Morales Maqueda et al., 2004]. Water column properties in the case of sensible heat polynyas are largely determined by cooling as opposed to net ice production and subsequent salinization in latent heat polynyas [Smith et al., 1990].

These two main mechanisms of polynya formation are not mutually exclusive and many times both contribute to the maintenance of the polynya, although one is typically dominant [Smith et al., 1990]. In addition, sea ice melting driven by solar radiation in spring and summer governs polynya maintenance, rendering the distinction between sensible and latent heat polynyas irrelevant [Morales Maqueda et al., 2004].

The Ross Sea polynya, which forms to the north of the Ross Ice Shelf on the southwestern Ross Sea (Figure 1), is the largest polynya in the region which has on the average 27,000 km² and a maximum of 50,000 km² open water area [Zwally et al., 1985]. Sea ice cover of the Ross Sea polynya area can be described as “open ice” in winter [World Meteorological Organization, 1970]. The polynya initially forms between the end of October and late November, on the average on November 7 [Arrigo et al., 1998]. The Ross Sea polynya rapidly expands in size in November and is the initiation site for sea ice melting in spring [Smith, 1995; Arrigo et al., 1998]. The polynya edge extends along the ice shelf with maximum width attained over western part of the continental shelf [Morales Maqueda et al., 2004]. Sea ice thickness to the south of 74°S in this region ranges between 0.1-0.7 m in austral fall and winter [Jeffries and Adolphs, 1997; DeLiberty and Geiger, 2005] and 0.1-0.4 m in spring [Arrigo et al., 2003].
The formation of the Ross Sea polynya is wind-driven [Zwally et al., 1985; Jacobs and Comiso, 1989], arguably linked to katabatic wind intensification [Bromwich et al., 1998] or strongly influenced by synoptic winds [Zwally et al., 1985]. North-westward katabatic airflow, with velocities higher than 30 m s\(^{-1}\), are estimated to drive 60\% of polynya events during winters between 1988 and 1991 [Bromwich et al., 1998]. Sensible heat supplied by warm water intrusions onto the shelf contribute to maintenance of the open water area in austral spring and summer [Jacobs and Comiso, 1989; Jacobs et al., 1995]. Sea ice cover starts to decay from south to north in early spring and polynya extent reaches the ice margin by January [Morales Maqueda et al., 2004]. While heat input from the ocean and wind forcing likely contribute to the increase in polynya size, rapid expansion of the polynya is mostly governed by increased solar radiation absorption by the ocean surface [Fichefet and Goosse, 1999].

Modeling studies of the Ross Sea polynya area [Fichefet and Goosse, 1999; Assmann et al., 2003; Reddy et al., 2007] suggested that the polynya is wind-driven. Reddy et al. [2007] showed that neither the non-advecting sea ice nor the zero-winds versions of coupled sea ice-ocean simulations yielded the summer Ross Sea polynya. Studies also suggest that interannual variability in the timing [Arrigo et al., 1998] and the size [Bromwich et al., 1998] of the polynya are controlled by winter temperatures, and that heat entrainment to the ocean surface layer from the underlying warm Circumpolar Deep Water contributes to the maintenance of open water fraction [Jacobs and Comiso, 1989].

Terra Nova Bay polynya (Figure 1) is a smaller feature, on the average 1,000 km\(^2\) of open water ranging between 0 and 5000 km\(^2\), and is formed and maintained by persistent westerly katabatic winds which average 13 m s\(^{-1}\) in speed and are stable tens of kilometers offshore [Bromwich and Kurtz, 1984; Kurtz and Bromwich, 1983]. The polynya is oriented in the north-south direction and is located between the land and the Drygalski Ice Tongue. Winds blowing from the west advect sea ice out of the polynya region and the ice tongue prevents northward advection of sea ice into the bay [Bromwich and Kurtz, 1984; Kurtz and Bromwich, 1983]. Surface heat budget calculations predict that Terra Nova Bay polynya may produce the equivalent of 10\% of the total sea ice production over the Ross Sea continental shelf, which corresponds to a cumulative annual sea ice production of 60 m [Kurtz and Bromwich, 1985].

Terra Nova Bay polynya is mainly a latent heat feature; however, Van Woert [1999a] shows sensible and longwave heat fluxes explain up to 50\% of the variance in the open water extent. The polynya plays an important role in the maintenance of High Salinity Shelf Water (HSSW) by mixing 60\% of the high density brine with Low
Salinity Shelf Water (LSSW) while the remainder is carried away in surface layers [Kurtz and Bromwich, 1985]. Dense water formation estimated by the conversion of LSSW to HSSW corresponds to a production of approximately 1 Sv ($10^6$ m$^3$s$^{-1}$) [Van Woert, 1999a].

The second part of this study will focus on the following research questions in relation to polynya dynamics: What are the processes responsible for the formation and maintenance of polynyas in the Ross Sea? What are the relative contributions of sensible and latent heat to ocean surface heat flux in the polynya areas? What is the role of katabatic winds on coastal polynya formation?

Following the small and basin scale investigation of sea ice dynamics, the third part of the study will focus on the effects of sea ice dynamics on ocean circulation, hydrography, and water mass transformation with the research questions: What are effects of the sea ice, polynya processes and associated surface heat, salt, and momentum fluxes on the water column structure? How are the resultant water masses distributed, transformed, and transported?

I.3 SEA ICE-OCEAN MODELING

Sea ice models are designed to simulate the sea ice mass balance, which is a product of its area and thickness, achieved by thermodynamics (formation and melting) and redistributed by dynamics (drift, ridging, and lead opening). Polar sea ice packs consist of open water and ice in different thickness ranges, i.e. thin first-year ice, thicker multiyear ice, and thick ice in the form of pressure ridges. Many thermodynamic and dynamic properties of the sea ice pack, such as the compressive strength, surface temperature, turbulent, radiative, and conductive flux exchange with the atmosphere, and rate of growth/melt, depend on ice thickness [Thorndike et al., 1975]. While dynamics drive sea ice drift and deformation (lead opening and ridging), thermodynamics respond with accretion and ablation to regulate the sea ice thickness distribution. Thus the sea ice model describes the evolution of the ice thickness distribution (ITD) in time and space as determined by the competition between these two processes [Briegleb et al., 2004].

Coupled sea ice-ocean models have been implemented to study large scale sea ice-ocean dynamics, as well as, smaller scale polynya processes. Large scale studies for the Ross Sea area consist of low resolution models [Stössel et al., 1990; Fichefet and Goosse, 1999] of sea ice-ocean dynamics and three-dimensional modeling studies of atmosphere coupled with a frazil and consolidated sea ice model [Gallée, 1997]. Sea ice
dynamics at the surface ocean are studied by one-dimensional coupled sea ice-mixed layer-thermocline models [Owens and Lemke, 1990; Smith and Klinck, 2002]. One-dimensional models [Pease, 1987; Darby et al., 1995; Van Woert, 1999a,b; Tear et al., 2003] are used to model the polynya width as determined by the balance between ice production and the flux of ice out of the polynya region. Ocean circulation [Dinniman et al., 2003, 2007] and mixed layer [Markus, 1999] models that do not have a dynamic sea ice component use observed ice concentration as forcing.

Sea ice models, like other components of earth system modeling, are subject to limitations of spatial and temporal resolution, and the results are affected by error sources such as parameterizations of the modeled processes and inaccurate forcing fields [Timmermann et al., 2002; Stössel and Markus, 2004; Lindsay and Zhang, 2006].

I.3.1 Data for Sea Ice Model Validation

Satellite passive microwave radiometers provide the most spatially and temporally comprehensive global sea ice concentration data [Comiso et al., 1997], widely and consistently used for observational studies [e.g., Sturman and Anderson, 1986; Jacobs and Comiso, 1989; Bromwich et al., 1998; Budillon et al., 2000; Kwok, 2005] and model validation [e.g., Fichefet and Morales Maqueda, 1997; Weatherly et al., 1998; Kreyscher et al., 2000; Timmermann et al., 2002; Budgell, 2005]. Therefore, among the most commonly modeled sea ice state variables (i.e. concentration, thickness, internal stress, surface temperature, velocity, enthalpy, snow thickness), sea ice concentration and velocity are the most directly validated due to the extensive spatial coverage and high frequency of continuous Defense Meteorological Satellites Program (DMSP) Special Sensor Microwave/Imager (SSM/I) data record. The National Aeronautics and Space Administration (NASA) Team [Cavalieri et al., 1990] and the Bootstrap [Comiso, 1995] algorithms are developed techniques to derive sea ice concentrations from multichannel data. Comiso et al. [1997], in a study comparing two techniques, reported that in some areas in the Antarctic, sea ice concentration values obtained by the Bootstrap algorithm may be as much as 25% higher and 30% lower than those obtained by the NASA Team algorithm, while available Landsat, advanced very high resolution radiometer (AVHRR), and synthetic aperture radar (SAR) datasets all yield higher concentrations compared to either of the passive microwave algorithms. For the Ross Sea in particular, the difference between the Bootstrap and NASA Team monthly mean values range between 20% and -15% [Comiso et al., 1997]. The accuracy of the total NASA Team sea ice concentration data is reported to be within ±5% in winter and ±15% during melting season [Cavalieri et al., 1996]. Accuracy
of the data increases within the consolidated ice pack when the concentration is high and sea ice is thicker than 0.2 m, and decreases as the fraction of thin ice increases [Cavalieri et al., 1996] and at low sea ice concentrations [Zibordi et al., 1995]. For the Bootstrap field, accuracy is estimated to be 5-10% in general and is reported to decrease within the pixel in the presence of high proportions of thin sea ice, melt ponds, or wet snow due to submerged freeboard [Comiso, 1999].

Limited data are available for a complete validation of the seasonal sea ice thickness distribution. Two studies provide characteristics of the climatological circumpolar distribution and seasonal variations: data reported by Timmermann et al. [2004] include the mean circumpolar field from the Antarctic Sea Ice Processes and Climate (ASPeCt) dataset by Worby and Ackley [2000] (Figure 2a) and its form corrected for sampling bias (Figure 2b). DeLiberty and Geiger [2005] provide the distribution for the Ross Sea from four weeks of National Ice Center (NIC) analysis (Figure 3).
Figure 3. Sea ice thickness in the Ross Sea based on NIC analysis. The weeks shown are (a) 1-8 June 1995, (b) 30 May-5 June 1998, (c) 23-29 January 1999, and (d) 22-28 January 2000. Figure reproduced from DeLiberty and Geiger [2005].
CHAPTER II
METHODS

The research questions raised in Chapter I are addressed using a coupled sea ice-ocean model. Theoretical and numerical aspects of the dynamic-thermodynamic sea ice component of the coupled model are presented in Section II.1. The hydrostatic, primitive equation ocean general circulation model that includes ice shelf cavity processes is described in Section II.2. Section II.3 provides the details of one- and three-dimensional coupled model implementation.

II.1 THE SEA ICE MODEL

The sea ice model used in this study is the Community Ice Code (CICE), version 3.1 [Hunke and Lipscomb, 2004]. CICE has been developed at the Los Alamos National Laboratory (LANL) as a component of a global climate model and is compatible with the Parallel Ocean Program (POP) [Smith et al., 1992] of LANL. CICE is also in conformance with the Community Sea Ice Model, version 5 (CSIM5) [Briegleb et al., 2004], the sea ice component of the Community Climate System Model, version 3.0 (CCSM3) [Kiehl and Gent, 2004], of the National Center for Atmospheric Research (NCAR). CICE is vectorized and uses two-dimensional domain decomposition and time-split thermodynamics-dynamics for efficient parallel computing in distributed memory platforms.

The major components of CICE are models of sea ice thermodynamics, dynamics, and horizontal transport that interactively describe the state and motion of sea ice. Model formulation and the numerical solution are presented in the sections that follow: Section II.1.1 provides an overview of CICE and introduces the state variables and fundamental equations. The horizontal transport scheme for sea ice advection is discussed in Section II.1.2. The thermodynamic model and related transport in thickness space are described in Section II.1.3. Sea ice dynamics and mechanical redistribution in thickness space due to ridging and rafting processes are outlined in Section II.1.4. Section II.1.5 provides the details of the numerical implementation.

II.1.1 Model Formulation

CICE consists of an energy-conserving thermodynamics model [Bitz and Lipscomb, 1999], featuring a resolved vertical temperature profile and an explicit brine pocket
parameterization, for the calculation of the local growth rates of snow and sea ice; a model of elastic-visco-plastic (EVP) sea ice dynamics [Hunke and Dukowicz, 1997], which includes the effects of metric terms [Hunke and Dukowicz, 2002], for the computation of the velocity field of the ice pack based on a parameterization of the material strength of sea ice [Rothrock, 1975]; a transport model that defines the horizontal advection of ice state variables by a second-order accurate, incremental remapping scheme [Lipscomb and Hunke, 2004]; and an energy-based ridging parameterization that transfers sea ice among thickness categories [Thorndike et al., 1975; Hibler, 1980]. Sea ice thickness distribution is Lagrangian [Thorndike et al., 1975; Bitz et al., 2001] and thickness space evolution due to thermodynamics is linearly remapped [Lipscomb, 2001]. Lateral and bottom melting are formulated as in McPhee [1992].

Sea ice is partitioned into discrete thickness categories. Each thickness category is represented by one or more layers of ice and one layer of snow. Prognostic variables for each thickness category are ice area fraction, ice volume, ice internal energy in each vertical layer, snow volume, and snow/ice surface temperature. Flux exchange with the atmosphere and the ocean is evaluated over each thickness category and aggregated. A nonlinear, vertical salinity profile for each thickness category remains constant. The longwave radiation and latent and sensible heat fluxes depend on the snow/ice surface temperature, which is calculated by a nonlinear flux balance. The albedo parameterization, which assumes a prescribed fraction of melt ponds, depends on surface type, surface temperature, and both the ice and snow thicknesses. Snow/ice albedo values are aggregated based on snow "patchiness". Prognostic variables ice velocity and the associated stress tensor are not resolved across the thickness distribution.

**Governing equations**

The governing equation used in CICE [Thorndike et al., 1975] to describe the evolution of the ice thickness distribution (ITD) in time \( t \) and space \( x \) is:

\[
\frac{\partial g}{\partial t} = -\nabla \cdot (gu_i) - \frac{\partial}{\partial h} (fg) + \psi, 
\]

where \( g = g(x, h, t) \) is the ice thickness distribution function, i.e. the fractional area covered by ice in the thickness range \((h, h + dh)\), \( u_i \) is the horizontal velocity, \( \nabla = (\partial/\partial x, \partial/\partial y) \), \( f \) is the rate of thermodynamic growth, and \( \psi \) is a redistribution function due to rafting and ridging. The cumulative distribution function is \( G(h) = \int_0^h g(h)dh \), where \( \int_0^\infty g(h)dh = 1 \). The aggregate ice fraction is
Table 1. Description and units of state variables for the sea ice model

<table>
<thead>
<tr>
<th>State variable</th>
<th>Description</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a_n$</td>
<td>Sea ice area</td>
<td>fraction</td>
</tr>
<tr>
<td>$v_{nl}$</td>
<td>Sea ice volume</td>
<td>m$^3$ m$^{-2}$</td>
</tr>
<tr>
<td>$e_{nl}$</td>
<td>Sea ice internal energy</td>
<td>J m$^{-2}$</td>
</tr>
<tr>
<td>$v_{sn}$</td>
<td>Snow volume</td>
<td>m$^3$ m$^{-2}$</td>
</tr>
<tr>
<td>$T_{sfn}$</td>
<td>Snow/ice surface temperature</td>
<td>°C</td>
</tr>
<tr>
<td>$u_i$</td>
<td>Sea ice velocity</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>$\sigma_{ij}$</td>
<td>Stress tensor components</td>
<td>N m$^{-1}$</td>
</tr>
</tbody>
</table>

$a_n = 0, 1, 2, \ldots, N$ denotes the category index. Category 0 is open water.

Equation (1) is solved by replacing the continuous function $g(h)$ by a discrete variable $a_n$, defined as the fractional area covered by ice in the thickness range $(H_{n-1}, H_n]$ for category indices $n = 1, 2, \ldots, N$. Category 0 denotes open water. Hence, by definition, $\sum_{n=0}^{N} a_n = 1$. The prognostic state variables of the sea ice model are listed in Table 1. Subscript $l = 1, 2, \ldots, L$ denotes the layer index. Sea ice and snow volumes are “per grid cell area” and thus have units of meters.

For each category, ice thickness is $h_n = \sum_{l=1}^{L} v_{nl} / a_n$ and lies within constant category limits. Equation (1) is integrated over the thickness limits for each category, resulting in a discrete set of $N$ equations to be solved for the sea ice fraction in each category $n$:

$$a_n = \int_{H_{n-1}}^{H_n} g(h) dh.$$  \hspace{1cm} (2)

Ice volume, $v_n = \int_{H_{n-1}}^{H_n} h g(h) dh$, is the first moment of (2). Any function $F$, which is linear in the ice thickness, therefore, is $F_n = F_0 a_n + F_1 v_n$.

The fundamental equation (1) applies to the state variables as:

$$\frac{\partial F_n}{\partial t} = S_{TF_n} - \nabla \cdot (F_n u_i) - S_{MF_n}, \quad n = 1, 2, \ldots, N,$$  \hspace{1cm} (3)

where $F_n$ stands for $a_n, v_n, e_n$, or $v_{sn}$, and $S_T$ and $S_M$ denote sources and sinks due
to thermodynamic processes and mechanical redistribution, respectively.

Snow thickness is $h_{sn} = v_{sn}/a_{n}$ and snow energy per unit volume is obtained from $e_{sn} = q_s v_{sn}$, where $q_s$ is the constant energy of melting (enthalpy) of snow. The internal sea ice energy is $e_n = q_n v_n$, where the proportionality function $q_n = q_n(T_n, S_n)$ (energy of melting, enthalpy) is the internal energy per unit volume. $q_n$ explicitly represents the effects of brine pockets and depends on temperature $T_n$ and salinity $S_n$. Using a prescribed salinity profile $S_n$, the vertical temperature profile is inferred by solving for $T_n$ in $q_n(T_n, S_n) = e_n/v_n$. The heat equation governing vertical heat transfer over time interval $t$ to $t'$ corresponding to temperatures $T_n$ and $T'_n$, respectively, is obtained over an ice thickness $h_n = v_n/a_{n}$ by

$$\int_{T_n}^{T'_n} \rho_i c_i dT_n = \int_t^{t'} \left( \frac{\partial}{\partial z} k \frac{\partial T_n}{\partial z} + Q_{SW} \right) dt, \quad n = 1, 2, \ldots, N, \quad (4)$$

where $\rho_i$ is the constant ice density (Table 2), $c_i$ is the temperature and salinity dependent ice heat capacity, $k$ is the snow or ice thermal conductivity, and $Q_{SW}$ is the absorbed shortwave radiation flux. The conservation of snow/ice surface temperature then is

$$\frac{\partial a_n T_{sf,n}}{\partial t} = S_{TT_{sf,n}} - \nabla \cdot (a_n T_{sf,n} u_i) - S_{MT_{sf,n}}, \quad n = 1, 2, \ldots, N. \quad (5)$$

The dynamics component of CICE assumes sea ice is a two-dimensional continuum. Therefore, the sea ice velocity $u_i$ and the stress tensor $\sigma_{ij}$ are representative of the entire ice thickness distribution. The momentum conservation equation is

$$m \frac{\partial u_i}{\partial t} = \nabla \cdot \sigma + \tau_a + \tau_w - mf k \times u_i - mg_e \nabla \eta, \quad (6)$$

where $m = \rho_s \sum_{n=1}^{N} v_{sn} + \rho_i \sum_{n=1}^{N} v_n$ ($\rho_s$ is the density of snow, Table 2), $f$ is the Coriolis parameter, $k$ is the local vertical unit vector, $\tau_a$ and $\tau_w$ are air and ocean stresses, respectively, $g_e$ is the acceleration due to gravity (Table 2), $\eta$ is the sea surface height, and $\nabla \cdot \sigma$ is the force per unit area due to internal ice stress. The nonlinear advection terms are negligibly small when Equation (6) is scaled, and hence, are ignored.

The governing equation for the stress tensor is

$$\frac{\partial \sigma_{ij}}{\partial t} + \frac{\epsilon^2}{2T_{ew} \sigma_{ij}} + \frac{1 - \epsilon^2}{4T_{ew}} \sigma_{kk} \delta_{ij} = \frac{P}{2T_{ew} \Delta t} \epsilon_{ij} - \frac{P}{4T_{ew}} \delta_{ij}, \quad i, j = 1, 2, \quad (7)$$
Table 2. Definition and values of the parameters and physical constants used in the sea ice model

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\rho_i$</td>
<td>Density of ice</td>
<td>917 kg m$^{-3}$</td>
</tr>
<tr>
<td>$\rho_s$</td>
<td>Density of snow</td>
<td>330 kg m$^{-3}$</td>
</tr>
<tr>
<td>$g_e$</td>
<td>Acceleration of gravity</td>
<td>9.81 m s$^{-2}$</td>
</tr>
<tr>
<td>$S_{\text{max}}$</td>
<td>Maximum salinity at ice base</td>
<td>3.2</td>
</tr>
<tr>
<td>$\rho_a$</td>
<td>Density of dry air</td>
<td>1.3 kg m$^{-3}$</td>
</tr>
<tr>
<td>$c_a$</td>
<td>Specific heat of air</td>
<td>1005 J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$L_v$</td>
<td>Latent heat of vaporization</td>
<td>$2.501 \times 10^6$ J kg$^{-1}$</td>
</tr>
<tr>
<td>$L_f$</td>
<td>Latent heat of fusion</td>
<td>$3.34 \times 10^8$ J kg$^{-1}$</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>Stefan-Boltzmann constant</td>
<td>$5.67 \times 10^{-8}$ W m$^{-2}$ K$^{-4}$</td>
</tr>
<tr>
<td>$\epsilon$</td>
<td>Emissivity of snow/ice</td>
<td>0.95</td>
</tr>
<tr>
<td>$\alpha_{\text{iv}}$</td>
<td>Visible ice albedo</td>
<td>0.68</td>
</tr>
<tr>
<td>$\alpha_{\text{in}}$</td>
<td>Near-infrared ice albedo</td>
<td>0.30</td>
</tr>
<tr>
<td>$\alpha_{sv}$</td>
<td>Visible snow albedo</td>
<td>0.96</td>
</tr>
<tr>
<td>$\alpha_{sn}$</td>
<td>Near-infrared snow albedo</td>
<td>0.68</td>
</tr>
<tr>
<td>$\alpha_w$</td>
<td>Ocean albedo</td>
<td>0.06</td>
</tr>
<tr>
<td>$\rho_w$</td>
<td>Density of seawater</td>
<td>1025 kg m$^{-3}$</td>
</tr>
<tr>
<td>$c_w$</td>
<td>Specific heat of seawater</td>
<td>4218 J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$c_{\text{dw}}$</td>
<td>Sea ice-ocean drag coefficient</td>
<td>0.00536</td>
</tr>
<tr>
<td>$c_0$</td>
<td>Specific heat of fresh ice at 0°C</td>
<td>2106 J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$\mu$</td>
<td>Ocean freezing temperature constant</td>
<td>0.0544°C</td>
</tr>
<tr>
<td>$k_0$</td>
<td>Thermal conductivity of fresh ice</td>
<td>2.03 W m$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$\beta$</td>
<td>Thermal conductivity ice constant</td>
<td>0.1172 W m$^{-1}$</td>
</tr>
<tr>
<td>$k_s$</td>
<td>Thermal conductivity of snow</td>
<td>0.3 W m$^{-1}$ K$^{-1}$</td>
</tr>
</tbody>
</table>

where $(i, j = 1, 2)$ represent the four components of the stress tensor, $e$ is a constant ratio of major to minor axes of the elliptical yield curve, $T_{ew}$ is a damping time scale for elastic waves, $\delta_{ij}$ is the Kronecker delta, $P$ is the ice compressive strength (i.e. mechanical pressure) as a function of the thickness distribution, $\dot{e}_{ij}$ is the rate of strain tensor, and $\Lambda'$ is a function of the rate of strain tensor.

The three terms on the right-hand side of Equation (1) describe three different kinds of sea ice transport: (i) horizontal transport in $x = (x, y)$ space, (ii) transport in thickness space $h$ due to thermodynamic processes, and (iii) redistribution in thickness space $h$ due to dynamics, ridging, and other mechanical processes. The equation is solved by operator splitting in three stages, where at each stage all but one of the three terms on the right is set to zero.
II.1.2 Horizontal Transport

The continuity equation

$$\frac{\partial F_n}{\partial t} + \nabla \cdot (F_n \mathbf{u}_i) = 0,$$

obtained by setting the second and third terms on the right-hand side of (1) to zero, describes the conservation of a state variable under horizontal transport for each ice thickness category $n$. The equation applies to fractional ice area ($a_n$) and tracers, namely ice volume ($v_n$), ice energy ($e_n$), snow volume ($v_{sn}$), and area-weighted snow/ice surface temperature ($a_n T_{sfn}$). Ice and snow densities are kept constant and hence volume conservation is equivalent to mass conservation. Fractional area of open water ($n = 0$) is also transported.

CICE features two transport schemes, multidimensional positive definite advection transport algorithm (MPDATA) [Smolarkiewicz, 1984] and incremental remapping [Dukowicz and Baumgardner, 2000] as modified for sea ice [Lipscomb and Hunke, 2004]. In this study, incremental remapping is used because it is less diffusive than the upwind scheme and is computationally faster than MPDATA [Lipscomb and Hunke, 2004]. Incremental remapping conserves the quantity (area, volume, or energy) being transported, is non-oscillatory (i.e. it does not create unphysical ripples in the transported fields), and preserves monotonicity (i.e. it does not create false extrema). The scheme is second-order accurate in space when not reduced locally to first-order to preserve monotonicity and is efficient for large numbers of categories and tracers, because much of the computation, which is geometrical, is performed only once per grid cell rather than per quantity transported.

Calculation of horizontal transport using the incremental remapping algorithm proceeds in three steps, (1) reconstructing area and tracer fields, (2) identifying departure triangle locations, and (3) transferring flux and updating state variables, which are described as follows.

(1) Area and tracer field reconstruction

Using the values of the state variable, area and tracer fields are reconstructed in each grid cell as linear functions of the position vector $\mathbf{r} = (x, y)$ relative to the cell center. For each field the value at the cell center and the gradients in the $x$ and $y$ directions are computed. Monotonicity is ensured by limiting the gradients, following van Leer [1979], cited in Dukowicz and Baumgardner [2000], such that the reconstructed field values lie within the range of the mean values in the grid cell and its eight neighbors. When integrated over the grid cell, the means of the reconstructed
fields are equal to the values of the state variables, denoted by $\bar{a}$, $\bar{h}$, and $\bar{q}$ for ice area, thickness, and enthalpy, respectively. Considering the fractional ice area for instance, the reconstructed field $a(r)$ should have the form

$$a(r) = \bar{a} + \alpha_a \langle \nabla a \rangle \cdot (r - \bar{r}),$$  \hspace{1cm} (9)

where $\langle \nabla a \rangle$ is the centered estimate of the area gradient within the cell, $\alpha_a$ is the limiting coefficient, and $\bar{r} = \frac{1}{A} \int_A r \, dA$ is the cell centroid, to satisfy

$$\int_A a \, dA = \bar{a}A,$$  \hspace{1cm} (10)

where $A = \int_A dA$ is the grid cell area. The reconstructed ice thickness $h(r)$ and enthalpy $q(r)$ are given by

$$h(r) = \bar{h} + \alpha_h \langle \nabla h \rangle \cdot (r - \bar{r}),$$  \hspace{1cm} (11)

$$q(r) = \bar{q} + \alpha_q \langle \nabla q \rangle \cdot (r - \bar{r}),$$  \hspace{1cm} (12)

where $\alpha_h$ and $\alpha_q$ are limiting coefficients, and

$$\bar{r} = \frac{1}{\bar{a}A} \int_A ar \, dA \quad \text{and} \quad \bar{r} = \frac{1}{\bar{a}hA} \int_A ah \, dA,$$  \hspace{1cm} (13)

are the centroids of ice thickness and enthalpy, respectively. The reconstructed ice thickness ($h(r)$) and enthalpy ($q(r)$) then satisfy

$$\int_A ah \, dA = \bar{a}hA,$$  \hspace{1cm} (14)

$$\int_A ahq \, dA = \bar{a}h\bar{q}A.$$  \hspace{1cm} (15)

Snow thickness and the surface temperature are treated like ice thickness.

(2) Departure triangle locations

Using the ice velocities at grid cell corners, the departure regions for fluxes across each grid cell edge are identified, these regions are divided into triangles, and the coordinates of the triangle vertices are computed. First, four cell corner velocities are projected directly backwards and a quadrilateral departure region is formed by connecting the starting points of these vectors (Figure 4). The location of a vertex
Figure 4. An illustration of departure triangles in incremental remapping. Shaded departure region is formed by connecting the starting points of the backwards trajectories of velocity vectors from four cell corners. Conserved quantities are remapped from the departure region to the "home cell" H across its edges shared with the north (N) and east (E) neighbor cells. The regions fluxed from the north edge are the triangles abc, acd, and ade. W, NW, and NE denote the west, northwest, and northeast neighbor cells, respectively. Figure reproduced from Lipscomb and Hunke [2004].

of the departure region (i.e. departure point), $x_D$, is defined as

$$x_D = u'_j A \Delta t,$$

(16)

where $u'$ denotes the velocity vector defined in the cell-corner basis. Fluxes are computed across the north and east edges of each grid cell for the entire domain. Considering the example in Figure 4, across the north edge of the home cell there are fluxes from the neighboring NW and N cells. H receives the triangles acd and ade from N and abc from NW. The coordinates of the vertices are determined using Euclidean geometry.

(3) Flux transfer and updating state variables

Each field is integrated over the triangles to obtain fluxes of area, volume, and energy. The fluxes are then transferred across cell edges and the state variables are updated. The integral $I_i$ of a linear function $f(r)$ over a triangle with vertices
\( \mathbf{x}_i = (x_i, y_i), i = 1, 2, 3, \) is given by

\[
I_i = A_T f(x_0),
\]

where

\[
\mathbf{x}_0 = (x_0, y_0) = \frac{1}{3} \sum_{i=1}^{3} x_i
\]

is the triangle midpoint and

\[
A_T = \frac{1}{2} \left( (x_2 - x_1)(y_3 - y_1) - (y_2 - y_1)(x_3 - x_1) \right)
\]

is the triangle area. Using (9), fractional ice area at the midpoint is given by

\[
a(x_0) = a_c + a_x x_0 + a_y y_0,
\]

where \( a_x = \alpha \frac{\partial a}{\partial x} \) and \( a_y = \alpha \frac{\partial a}{\partial y} \).

It follows from (14) and (15) that the ice volume and energy fluxes are evaluated using the integrals of a quadratic and a cubic function, respectively. The integral \( I_q \) of a quadratic polynomial over a triangle is given by a three-point formula:

\[
I_q = \frac{A_T}{3} \sum_{i=1}^{3} f(x_i'),
\]

where \( x_i' = (x_0 + x_i)/2 \). The integral \( I_c \) of a cubic polynomial, on the other hand, is evaluated using a four-point formula:

\[
I_c = A_T \left[ -\frac{9}{16} f(x_0) + \frac{25}{48} \sum_{i=1}^{3} f(x_i'') \right],
\]

where \( x_i'' = (3x_0 + 2x_i)/5 \). Calculated fluxes are used to update the values of the state variable for each thickness category. The updated fractional ice area \( a_{\text{new}} \) for grid cell \((i, j)\) for instance is

\[
a_{\text{new}} = a(i, j) + \frac{1}{A(i, j)} \left[ F_E(i-1, j) - F_E(i, j) + F_N(i, j-1) - F_N(i, j) \right],
\]

where \( F_E(i, j) \) and \( F_N(i, j) \) are area fluxes across the east and north edges, respectively, and \( A(i, j) \) is the grid cell area. Conservation of a transported global quantity is maintained by subtracting from the neighbor cell the flux added to the home cell.
The ice volume and energy as well as the snow volume and area-weighted surface temperature are updated analogously.

Steps 1 and 3 are repeated for each field in each ice thickness category. Step 2 is done only once per time step, because all fields are transported by the same velocity field. Remapping assumes that the velocities are defined at cell corners and scalars at cell centers, and therefore is intrinsically a B-grid scheme \cite{Arakawa1977}. Time-stepping is first-order accurate. Equation (8) is solved for velocities at time $t$ using the values of the state variables at time $t$ to remap area and tracers to time $t + \Delta t$.

II.1.3 Thermodynamics and Transport in Thickness Space

Setting the first and third terms on the right-hand side of (1) to zero yields the equation for ice transport in thickness space ($h$) due to thermodynamic growth and melting:

$$\frac{\partial g}{\partial t} + \frac{\partial}{\partial h}(fg) = 0. \quad (24)$$

The thermodynamic sea ice model that determines the rate of growth/melt ($f$) is based on \textit{Maykut and Untersteiner} \cite{Maykut1971} and \textit{Bitz and Lipscomb} \cite{Bitz1999}. First, the temperature profile and thickness changes in ice and snow are determined by an energy balance of radiative, turbulent, and conductive heat fluxes and a mass balance of evaporative flux and precipitation. Ice is then transported in thickness space by solving (24) using linear remapping method of \textit{Lipscomb} \cite{Lipscomb2001}, which is similar to the one-dimensional version of incremental remapping described in Section II.1.2.

\textbf{Thermodynamic sea ice model}

Sea ice thickness categories are treated as horizontally uniform columns divided into layers of equal thickness. Ice/snow surface temperature $T_{sf}$ has an upper limit of 0°C and the temperature at the bottom of the ice is held at $T_f$, the freezing temperature of the ocean surface layer. The snow is assumed to be fresh and a constant vertical salinity profile is prescribed for ice. Midpoint salinity ($S_i$) in each ice layer is given by

$$S_i = \frac{1}{2} S_{\text{max}} \left[ 1 - \cos \left( \pi z^{zb} \right) \right], \quad (25)$$

where $S_{\text{max}}$ is the maximum salinity at ice base (Table 2), $z = (l - 1/2)/L$, $a = 0.407$, and $b = 0.573$. The profile represents multiyear ice, where the salinity of the top layers are reduced by surface melt and has an average value of 2.3. First-year ice is vertically more uniform in salinity, however, since the effects of salinity on heat
capacity are small for below freezing temperatures, the resulting error in temperature is not significant.

Ice enthalpy \( q \), defined as the negative of the energy required to melt a unit volume of ice and raise its temperature to 0°C, depends on the brine pocket volume due to internal freezing and melting in brine pockets. Hence, ice enthalpy is a function of temperature and salinity. Since salinity is prescribed, there is a one-to-one relationship between ice enthalpy and temperature. Snow enthalpy \( q_s \) depends on temperature alone.

Ice and snow thicknesses, temperatures, and enthalpies at time \( t \) are advanced to time \( t + \Delta t \) by (1) calculating top and bottom surface thermodynamic forcing, (2) solving for new temperatures, and (3) computing growth and melt, as follows.

(1) Top and bottom surface thermodynamic forcing

The net atmosphere-ice energy flux (positive downward) is

\[
F_{top} = F_s + F_l + F_{L1} + F_{L1} + (1 - \alpha)(1 - i_0)F_{sw},
\]

where \( F_s, F_l, F_{L1}, F_{L1} \), and \( F_{sw} \) are the sensible, latent, incoming longwave, outgoing longwave, and incoming shortwave fluxes, respectively, \( \alpha \) is the shortwave albedo, and \( i_0 \) is the fraction of absorbed shortwave flux that penetrates into the ice. Turbulent heat fluxes and outgoing longwave radiation are functions of snow/ice surface temperature \( T_{sf} \), whereas incoming shortwave and longwave radiation are obtained from the atmosphere. The sensible and latent heat fluxes are proportional to the difference between air potential temperature, \( \Theta_a \), and \( T_{sf} \), and between specific humidity, \( Q_A \), and the surface saturation specific humidity, \( Q_{sf} \), respectively. For each category, sensible and latent heat fluxes, along with wind stress, are based on Monin-Obukhov similarity theory and are obtained from the bulk formulas

\[
F_s = C_s(\Theta_a - T^K_{sf}),\quad (27)
\]
\[
F_l = C_l(Q_a - Q_{sf}),\quad (28)
\]

where \( C_s \) and \( C_l \) are nonlinear turbulent heat transfer coefficients and \( T^K_{sf} \) is the surface temperature in Kelvin. The turbulent heat transfer coefficients are given by

\[
C_s = \rho_a c_a c_g |u_a| + 1,\quad (29)
\]
\[
C_l = \rho_a (L_v - L_f)c_q |u_a|,\quad (30)
\]
where $\rho_a$ is the air density (Table 2), $c_a$ is the specific heat of air (Table 2), $L_v$ and $L_f$ are the latent heats of vaporization and fusion (Table 2), $u_a$ is the wind velocity, and $c_u$, $c_q$, and $c_q$ are turbulent exchange coefficients for momentum, sensible and latent heat, respectively. The evaporative flux to the atmosphere is given in terms of the latent heat as $F_{\text{evap}} = F_L/(L_v + L_f)$. The momentum exchange coefficient is used in the calculation of wind stress on ice, $\tau_a$, as,

$$\tau_a = \rho_a c_u^2 |u_a| u_a.$$

Incoming longwave radiation is calculated following Parkinson and Washington [1979],

$$F_{L1} = \sigma T_a^4 (1 + 0.275c) \left\{ 1 - 0.261 \exp \left[ -7.77 \times 10^{-4} (273.16 - T_a)^2 \right] \right\},$$

where $\sigma$ is the Stefan-Boltzmann constant (Table 2), $T_a$ is the air temperature, and $c$ is the cloud fraction. Outgoing longwave radiation, $F_{L1}$, is assumed to follow blackbody radiation given by

$$F_{L1} = \epsilon \sigma (T_{sf}^K)^4,$$

where $\epsilon$ is the emissivity of snow/ice (Table 2).

Total shortwave radiation is divided into three parts: (i) a reflected portion, $-\alpha F_{sw}$, (ii) a portion that penetrates into the ice interior, $i_0(1-\alpha)F_{sw}$, and (iii) the remainder that is absorbed at the snow/ice surface.

The net absorbed shortwave flux, $F_{\text{swabs}} = (1 - \alpha)F_{sw} = \sum (1 - \alpha_j)F_{sw}^j$, where the summation is over four radiative categories: the direct and diffuse components of visible (< 700 nm) and near-infrared (> 700 nm) spectra (Table 2). The snow/ice albedo depends on the the snow/ice thickness, surface temperature, and the spectral band. The albedo parameterization ignores the zenith angle dependence and horizontal variations in snow/ice topography while including the effects of a prescribed melt pond fraction. The asymptotic visible and near-infrared albedo values $\alpha_{iv}$ and $\alpha_{in}$ for non-melting bare sea ice thicker than 0.5 m approximately fit the data of Allison et al. [1993] and Ebert and Curry [1993]. The sea ice albedo gradually decreases to the ocean albedo, $\alpha_w$, as the sea ice thickness goes to zero. Dry snow spectral albedos $\alpha_{sv}$ and $\alpha_{sn}$ are consistent with those of Grenfell et al. [1994] and Ebert and Curry [1993]. The directional and spectral albedos are weighted and merged into a single broad band value. The weights used in the model match those of the International Satellite Cloud Climatology Project (ISCCP) [Schiffer and Rossow, 1983] shortwave
forcing.

The flux penetrating into the ice is \( I_0 = i_0 F_{swabs} \), where \( i_0 = i_c[1 - h_s/(h_s + 0.02)] \) and \( i_c = 0.70 \) for visible and 0 for near-infrared radiation. The fraction \( (1 - i_0) \) is absorbed in an infinitesimal layer at the ice surface. Radiation penetrating into the ice attenuates according to Beer’s Law, \( I(z) = I_0 e^{-\kappa_i z} \), where \( I(z) \) is the shortwave flux that reaches depth \( z \) beneath the surface without being absorbed and \( \kappa_i \) is the bulk attenuation coefficient for solar radiation in ice. For category \( n \), a fraction \( e^{-\kappa_i h_n} \) of the penetrating shortwave radiation passes through the sea ice to the ocean.

The net ice-ocean heat flux (positive downward), based on boundary layer theory, is given by McPhee [1992]:

\[
F_{bot} = -\rho_w c_w c_h u_*(T_w - T_f),
\]

where \( \rho_w \) is the density of seawater (Table 2), \( c_w \) is the specific heat of seawater (Table 2), \( c_h = 0.006 \) is an empirical heat transfer coefficient, \( u_* = \sqrt{|\tau_w|/\rho_w} \) is the ice-ocean friction velocity used under melting conditions, \( T_w \) is the sea surface temperature, and \( T_f \) is the salinity-dependent ocean surface freezing point. The ice-ocean stress \( \tau_w \), used to calculate the friction velocity, is determined from the ocean surface current \( u_w \) in the form of a nonlinear drag law

\[
\tau_w = c_{dw}\rho_w |u_w - u_i|(u_w - u_i),
\]

where \( c_{dw} \) is the sea ice-ocean drag coefficient (Table 2).

(2) New temperatures

The temperature of each ice layer is coupled to the temperatures of the layers immediately above and below by heat conduction terms which are evaluated implicitly. The resulting set of equations are solved by a standard tridiagonal numerical integration algorithm. The rate of change of temperature in the ice interior is given by Maykut and Untersteiner [1971]:

\[
\rho_i c_i \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left( k_i \frac{\partial T}{\partial z} \right) - \frac{\partial I}{\partial z},
\]

where \( \rho_i \) is the sea ice density, \( c_i = c_i(T, S) \) is the specific heat of sea ice, and \( k_i = k_i(T, S) \) is the thermal conductivity of sea ice. The temperature \( T \) in °C and salinity \( S \) in psu are the ice temperature and salinity. The vertical coordinate \( z \) is defined as positive downward with \( z = 0 \) at the ice surface. The specific heat of
sea ice, which explicitly accounts for the storage of latent heat in brine pockets (i.e., inclusions of brine in sea ice), is originally given by Ono [1967], cited in Bitz and Lipscomb [1999]:

\[ c_i(T, S) = c_0 + \frac{L_f \mu S}{T^2}, \]  

(37)

where \( c_0 \) is the specific heat of fresh ice at 0°C and \( \mu \) is an empirical constant representing the ratio between the freezing temperature and salinity of brine (Table 2).

Thermal conductivity of sea ice, also influenced by brine pockets, is given following Untersteiner [1964] by Bitz and Lipscomb [1999]:

\[ k_i(T, S) = k_0 + \frac{\beta S}{T}, \]  

(38)

where \( k_0 \) is the conductivity of fresh ice and \( \beta \) is an empirical constant (Table 2).

The rate of change of temperature in snow is

\[ \frac{\rho_s c_s \partial T_s}{\partial t} = \frac{\partial}{\partial z} \left( k_s \frac{\partial T_s}{\partial z} \right), \]  

(39)

where \( T_s \) is the temperature, \( \rho_s \) is the density, \( c_s = c_0 \) is the specific heat, and \( k_s \) is the thermal conductivity of snow (Table 2). Most of the incoming sunlight is absorbed near the snow surface, which allows the penetrating solar radiation to be neglected.

(3) Growth and melting

For each ice category, \( n \), the ice enthalpy, \( q_n \), is evaluated using (37) by

\[ q_n(T, S) = -\rho_i \left[ c_0 (T_{\text{melt}} - T) + L_0 \left( 1 - \frac{T_{\text{melt}}}{T} \right) \right], \]  

(40)

where \( T_{\text{melt}} = -\mu S \) is the melting temperature of ice of salinity, \( S \). The enthalpy is defined to be negative. Hence, \( |q_n(T, S)| \) is the amount of heat that is required to raise the temperature of ice from \( T \) to \( T_{\text{melt}} \) at salinity \( S \), resulting in a rise of internal energy from \( e_n < 0 \) to zero. Equation (40) is quadratic in \( T \), and given the layer enthalpies, \( T \) corresponds to the root of

\[ \rho_i c_0 T^2 - (q_n + \rho_i c_0 T_{\text{melt}} + \rho_i L_f)T + \rho_i L_f T_{\text{melt}} = 0 \]  

(41)

that is below \( T_{\text{melt}} \).

The heat required to change snow temperature below melting is small compared to the latent heat of fusion, thus, ignoring that term, the constant enthalpy of snow is
given by

\[ q_s = -\rho_s L_f. \] (42)

The energy balance at the top and bottom surfaces determine the top melt and bottom growth and melt rates. Melting at the top surface is given by

\[
q \delta h = \begin{cases} 
(F_{\text{top}} + k \frac{\partial T_{sf}}{\partial z}) \Delta t & \text{if } (F_{\text{top}} + k \frac{\partial T_{sf}}{\partial z}) > 0 \\
0 & \text{otherwise}
\end{cases},
\] (43)

where \( q, k, \) and \( T_{sf} \) are the enthalpy, thermal conductivity, and temperature of the surface ice or snow layer, respectively, and \( \delta h \) is the change in thickness. Energy remaining, if any, after completely melting a layer is used to melt the layers beneath. Any flux left over once the ice and snow are completely melted is added to the ocean. Melt water is assumed to drain to the ocean.

Growth and melting at the bottom ice surface is given by

\[
q \delta h = - (F_{\text{bot}} + k_{\text{bot}} \frac{\partial T_{n}}{\partial z}) \Delta t,
\] (44)

where \( k_{\text{bot}} = k_i(T_f, S_{\text{max}}) \) is the thermal conductivity and \( T_n \) is the temperature of the bottom ice layer. For melting ice, \( q \) is the enthalpy of the bottom ice layer. Growing ice is added to the bottom layer and \( q \) is the enthalpy of the new ice with temperature \( T_f \) and salinity \( S_{\text{max}} \).

Atmospheric vapor is condensed at the surface as snow and ice if \( F_i > 0 \) and snow and ice sublimate if \( F_i < 0 \). Surface thickness change is given by

\[
(\rho L_v - q)\delta h = F_i \Delta t,
\] (45)

where \( \rho \) is the density and \( q \) is the enthalpy of the existing surface layer (snow or ice).

Snow to ice conversion occurs if the snow layer becomes thick enough to lie partially below the freeboard (the ocean surface). For snow of thickness \( h_s \) overlying sea ice of thickness \( h \), the snow-ice interface height, from Archimedes’ principle, is

\[
z_{\text{int}} = h - (\rho_s h_s + \rho_i h)/\rho_w.
\] (46)

For \( z_{\text{int}} < 0 \), an amount of snow equal to \(-z_{\text{int}} \rho_i/\rho_s\) is removed from the snow layer and is converted into ice of thickness \(-z_{\text{int}}\) with enthalpy \( q_s \rho_i/\rho_s\).

Lateral sea ice formation and melting depend on the sign of the freezing/melting potential, \( F_{\text{frzmlt}} \), of the underlying ocean (see Section II.3.1, Equation (115)). For
For \( F_{frazil} > 0 \), a volume \( v_f = F_{frazil} \Delta t/q_f \) of frazil ice is formed, where \( q_f = -\rho_i L_f \) is the enthalpy of frazil ice formation. Open water occurs for \( a_0 > 0 \) and newly formed frazil ice has thickness of \( h_f = v_f/a_0 \). The new ice is added to category 1 so long as \( h_f \) is less than the category thickness limit. New ice that exceeds this limit or \( a_0 = 0 \) results in the new ice being evenly distributed over all categories.

For \( F_{frazil} < 0 \), the melting potential is assumed to be dominated by solar radiation. Over a grid cell, solar radiation absorbed in the ocean surface layer where ice is thicker (thinner) than the grid cell mean ice thickness is available for lateral (basal) melting. The freezing/melting potential, \( F_{frazil} \), is partitioned into maximum available flux for lateral and basal melting by the fractions \( f_{lat} \) and \( f_{bot} \), respectively, according to

\[
\begin{align*}
    f_{bot} &= R \exp(-\bar{h}/\zeta_1) + (1 - R) \exp(-\bar{h}/\zeta_2), \quad (47) \\
    f_{lat} &= 1 - f_{bot}, \quad (48)
\end{align*}
\]

where \( \bar{h} \) is the grid cell mean ice thickness and \( R = 0.68, \zeta_1 = 1.2 \text{ m}^{-1}, \) and \( \zeta_2 = 28 \text{ m}^{-1} \) are empirical constants [Paulson and Simpson, 1977, cited in Briegleb et al. [2004]].

The actual amount of heat used for basal melting is \( F_{bot} \) and that for lateral melting is given by

\[
F_{lat} = E_{tot} M_a \frac{P_f}{A_f}, \quad (49)
\]

where \( E_{tot} \) is the vertically averaged aggregate snow and ice enthalpy, \( M_a \) is the interfacial melting rate [Maykut and Perovich, 1987], and \( P_f \) is the total floe perimeter per unit floe area \( A_f \) [Rothrock and Thorndike, 1984; Bitz et al., 2001]. The ice area, volume, energy, and snow volume are all reduced by the fraction \( R_{lat} = |F_{lat} \Delta t/E_{tot}| \) due to lateral melting.

The layer spacing changes after growth and melting, and therefore layer interfaces must be adjusted, conserving energy, to restore equal thickness. The overlap, \( \eta_{km} \), of each new layer, \( k \), with each old layer, \( m \), is given by

\[
\eta_{km} = \min(z_m, z_k) - \max(z_{m-1}, z_{k-1}), \quad (50)
\]

where \( z \) denotes the vertical coordinate of the corresponding layer. The adjusted enthalpies of new layers are

\[
q_k = \frac{h}{L} \sum_{m=1}^{L} \eta_{km} q_m, \quad (51)
\]
In addition to growth/melting, sublimation/condensation, and flooding, snowfall is added to the snow layer at the precipitation rate obtained from the atmosphere.

**Transport in thickness space by linear remapping**

The remapping method of Lipscomb [2001] is used for solving the conservation equation (24), in which the thickness categories are represented as Lagrangian volumes with time-dependent boundaries following the motion in thickness space. In each displaced category, a linear approximation to the thickness distribution function is remapped onto the original thickness categories.

For category $n$, the growth rate at $h = h_n$ is given by $f_n = (h_n^{t+\Delta t} - h_n^t)\Delta t$, where $h_n^t$ is the thickness of category $n$ at time $t$ and $h_n^{t+\Delta t}$ is the new thickness at time $t + \Delta t$, computed by the thermodynamic model. The time step, $\Delta t$, must be small enough that $h_n^{t+\Delta t} < h_{n+1}^{t+\Delta t}$ for $n = 1$ to $N - 1$. The growth rate at the upper category boundary, $H_n$, is

$$F_n = f_n + \frac{f_{n+1} - f_n}{h_{n+1} - h_n} (H_n - h_n).$$

(52)

The temporary displaced boundaries are given by

$$H_n^* = H_n + F_n \Delta t, \quad n = 1, \ldots, N - 1,$$

(53)

with the requirement that $H_{n-1} < H_n^* < H_{n+1}$. The ice areas in the displaced categories satisfy $a_n^{t+\Delta t} = a_n^t$, since ice area is conserved during vertical ice growth or melting. A positive and continuous linear approximation, $g(h)$, to the Lagrangian volume is constructed with the area and volume constraints given by

$$\int_{H_{n-1}^*}^{H_n^*} g(h) \, dh = a_n^m,$$

(54)

$$\int_{H_{n-1}^*}^{H_n^*} hg(h) \, dh = a_n^m h_n^{m+1} = v_n^{m+1}.$$

(55)

Given $g(h)$, the thickness distribution is remapped to the original boundaries by transferring area, $\Delta a_n$, and volume, $\Delta v_n$, between each original boundary, $H_n$, and displaced boundary, $H_n^*$. If $H_n^* > H_n$, ice moves from category $n$ to $n + 1$. The
transferred quantities are

$$\Delta a_n = \int_{H_n}^{H_n^*} g(h) \, dh,$$  \hspace{1cm} (56)

$$\Delta v_n = \int_{H_n}^{H_n^*} hg(h) \, dh.$$  \hspace{1cm} (57)

If $H_n^* < H_n$, the limits of integration of (56) and (57) are reversed and ice moves from category $n + 1$ to $n$. For the thinnest category ($n = 1$), if ice is growing in open water at a rate $F_0$, the boundary, $H_0$, is shifted to $F_0 \Delta t$ before $g(h)$ is constructed, then reset to zero after remapping. For no ice growth in open water, $H_0$ is fixed at zero and $F_0 = f_1$. The thickest category boundary, $H_N$, is not fixed and varies with $h_N$. Setting $H_N = 3h_N - 2H_{N-1}$ ensures positive $g(h)$ for $h < H_N$ and $g(0) = 0$ for $h \geq H_N$.

Transferred snow volume and ice energy in layer $l$ are $\Delta v_{sn} = v_{sn} (\Delta v_n / v_n)$ and $\Delta e_{nl} = e_{nl} (\Delta v_n / v_n)$, respectively. The snow/ice surface temperature changes with area due to transport in thickness space.

II.1.4 Dynamics and Mechanical Redistribution

Sea ice motion and deformation are governed by the wind stress, ocean drag, ocean surface slope, Coriolis force, and the internal ice stress. The internal stress tensor, $\sigma_{ij}$, in the momentum equation (6) gives the internal force on sea ice for a specified direction. The momentum equation in the CICE dynamics model provides ice velocity using the elastic-viscous-plastic (EVP) rheology [Hunke and Dukowicz, 1997], which is a modification of the standard viscous-plastic (VP) rheology of Hibler [1979]. Ice divergence or convergence due to deformation results in transport in thickness space, $h$, due to mechanical redistribution [Thorndike et al., 1975; Hibler, 1980], represented by $\psi$ in (1).

**Elastic-viscous-plastic rheology**

The EVP rheology relates the internal ice stress, $\sigma_{ij}$, and the rates of strain, $\dot{e}_{ij}$, using the gradients of the internal ice pressure, $P$, determined by an energy-based ridging scheme [Rothrock, 1975], such that the principal components of stress lie on an elliptical yield curve with $e = 2$. The general stress-strain relationship for fluids is given by

$$\sigma_{ij} = 2\eta\dot{e}_{ij} + (\zeta - \eta)\dot{e}_{kk}\delta_{ij} - \frac{P}{2}\delta_{ij},$$  \hspace{1cm} (58)
where $\dot{\epsilon}_{kk} = \dot{\epsilon}_{11} + \dot{\epsilon}_{22}$ is the total linear rate of strain (divergence), $\dot{\epsilon}_{12} = \dot{\epsilon}_{21}$ is the shear rate of strain, and $\zeta$ and $\eta$ are bulk (linear) and shear viscosities, respectively. The rate of strain tensor is a function of velocity gradients such that

$$\dot{\epsilon}_{ij} = \frac{1}{2} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right). \quad (59)$$

The ice strength is proportional to the change in ice potential energy per unit area of compressive deformation and is given by [Rothrock, 1975; Hibler, 1980]

$$P = \frac{1}{2} g_e \rho_i \left( \frac{\rho_w - \rho_i}{\rho_w} \right) C_f \int_0^\infty h^2 W_r \, dh, \quad (60)$$

where $C_f$ is an empirical parameter for frictional energy dissipation and $W_r$ is the ridging mode.

Equation (58) can be rewritten to represent the plastic contribution to the strain rate:

$$\dot{\epsilon}_{ij} = \frac{1}{2\eta} \sigma_{ij} + \frac{\eta - \zeta}{4\eta\zeta} \sigma_{kk} \delta_{ij} + \frac{P}{4\zeta} \delta_{ij}. \quad (61)$$

Defining divergence $D_D$, horizontal tension rate $D_T$, and shearing stress rate $D_S$ as $D_D = \dot{\epsilon}_{11} + \dot{\epsilon}_{22}$, $D_T = \dot{\epsilon}_{11} - \dot{\epsilon}_{22}$, and $D_S = 2\dot{\epsilon}_{12}$, respectively, $\Delta$ is given by

$$\Delta = \left[ D_D^2 + \frac{1}{e^2} (D_T^2 + D_S^2) \right]^{1/2}. \quad (62)$$

The viscosities $\zeta = P/2\Delta$ and $\eta = \zeta/\epsilon^2$ increase with pressure and decreasing strain rates.

The elastic part of the strain rate is approximated by

$$\dot{\epsilon}_{ij} = \frac{1}{E} \frac{\partial \sigma_{ij}}{\partial t}, \quad (63)$$

where $E$ is a parameter analogous to Young’s modulus.

Letting $\sigma_1 = \sigma_{11} + \sigma_{22}$ and $\sigma_2 = \sigma_{11} - \sigma_{22}$, the VP constitutive law is alternatively expressed by the stress equations

$$\frac{1}{E} \frac{\partial \sigma_1}{\partial t} + \frac{\sigma_1 + P}{2\zeta} = D_D, \quad (64)$$

$$\frac{1}{E} \frac{\partial \sigma_2}{\partial t} + \frac{\sigma_2}{2\eta} = D_T, \quad (65)$$

$$\frac{1}{E} \frac{\partial \sigma_{12}}{\partial t} + \frac{\sigma_{12}}{2\eta} = \frac{1}{2} D_S. \quad (66)$$
Equations (64-66) reduce to the VP rheology in steady-state or in the limit $E \to \infty$, while the elastic term (63) controls the solution and is a regularization of the VP rheology under conditions of very small strain rate, i.e. $\eta, \zeta \to \infty$. In contrast to the VP rheology, the stress tensor components are prognostic variables in the EVP model.

Defining the elastic parameter $E = \zeta / T_{ew}$ in terms of an elastic wave damping time-scale given by

$$T_{ew} = E_o \Delta t,$$

where $E_o < 1$ is a constant, Equations (64-66) become

$$\frac{\partial \sigma_1}{\partial t} + \frac{\sigma_1 + P}{2T_{ew}} = \frac{P}{2T_{ew} \Delta} D_D,$$

$$\frac{\partial \sigma_2}{\partial t} + \frac{\sigma_2}{2e^2T_{ew}} = \frac{P}{2T_{ew} \Delta} D_T,$$

$$\frac{\partial \sigma_{12}}{\partial t} + \frac{\sigma_{12}}{2e^2T_{ew}} = \frac{P}{4T_{ew} \Delta} D_S,$$

with all coefficients on the left-hand side constant except for $P$ (given by Equation 60). The momentum equation (6) and the EVP model stress tensor equations (68-70) are integrated over the subcycling elastic time step $\Delta t_e < T_{ew}$ for each time step $\Delta t$.

The left-hand side is treated implicitly and the right-hand side explicitly. The rate of strain $\dot{e}_{ij}$ and $\Delta$ are updated at each elastic time step, while $P$ changes on the longer time step, compensating for the reduced efficiency resulting from including the viscosities in the subcycling.

**Mechanical redistribution**

The term on the right-hand side of (1) describes the mechanical redistribution of ice in thickness space, $h$, due to rafting and ridging processes [Thorndike et al., 1975; Rothrock, 1975; Hibler, 1980; Flato and Hibler, 1995]. Applied after the horizontal transport calculation (Section (II.1.2), the mechanical redistribution scheme converts thinner ice to thicker, reducing the total ice area while conserving ice volume and energy and ensuring that the converging sea ice ridges enough so as to not exceed the grid cell area. The ridging of thin ice and closing of open water area over ridging of thicker ice is determined by a weighting function, $b(h)$, following Thorndike et al. [1975],

$$b(h) = \begin{cases} \frac{2}{C^*} \left(1 - \frac{G(h)}{G^*}\right) & \text{if } G(h) < G^* \\ 0 & \text{otherwise} \end{cases},$$

(71)
where $G(h)$ is the fractional area covered by ice thinner than $h$ and $G^* = 0.15$, an empirical constant, is the fractional area participating in ridging. The participation function, $a_P(h) = b(h)g(h)$, specifies the thickness distribution of the ice that participates in ridging. The mean value of $a_P$ in category $n$, $a_{P_n}$, represents the ratio of the ice area ridging and open water area closing in category $n$ to the total ice area ridging and open water area closing. The participation function ($a_{P_n}$) is obtained by integrating $a_P(h)$ between category boundaries $H_{n-1}$ and $H_n$ and is given by

$$a_{P_n} = \frac{2}{G^*}(G_n - G_{n-1}) \left(1 - \frac{G_{n-1} + G_n}{2G^*}\right).$$ (72)

For an open water fraction that is greater than the fractional area participating in ridging, $q_0 > G^*$, no ice can ridge which prevents reduction in the open water area. Setting $G^*$ too large causes thicker ice to participate in ridging, hence increasing the ice strength.

Ridging ice of thickness $h_n$ distributes ridge area uniformly between $H_n^{min} = 2h_n$ and $H_n^{max} = 2\sqrt{H^*h_n}$, where $H^* = 0.25$, an empirical constant, determines the mean thickness of ridged ice [Hibler, 1980]. The ratio of the mean ridge thickness to the thickness of ridging ice is $k_n = (H_n^{min} + H_n^{max})/2h_n$. If $r_n$ is the rate at which the fractional sea ice area is reduced by ridging, the areas of thicker categories grow at the rate $r_n/k_n$. The net rate of open water area closing combined with the net rate of ice area lost to ridging is expressed by $R_{net}$, following Flato and Hibler [1995], as

$$R_{net} = \frac{C_s}{2}(\Delta - |D_D|) - \min(D_D, 0),$$ (73)

where $C_s = 0.25$ is the fraction of shear dissipation energy that contributes to ridging. Divergence, $D_D$, and $\Delta$ are computed by the dynamics model. The area ridged in category $n$ is then given by $a_{rn} = r_n\Delta t$, where $r_n = a_{P_n}\sum_{n=0}^{N} r_n$. The area and volume of new ridges are $a_{rn}/k_n$ and $a_{rn}h_n$, respectively. Ridging ice from category $n$ is redistributed among the thicker categories. Also, when ice ridges, a fraction of the snow is assumed to fall to the ocean.

II.1.5 Numerical Solution

Five ice thickness categories and an open water category adequately resolve ice thickness, ice strength, and surface fluxes [Bitz et al., 2001; Lipscomb, 2001]. The thickness boundaries (Table 3) are based on the category limit formula [Lipscomb,
Table 3. Ice thickness category boundaries

<table>
<thead>
<tr>
<th>Category (n)</th>
<th>Thickness Range (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>1</td>
<td>0.0\textsuperscript{+}–0.65</td>
</tr>
<tr>
<td>2</td>
<td>0.65–1.39</td>
</tr>
<tr>
<td>3</td>
<td>1.39–2.47</td>
</tr>
<tr>
<td>4</td>
<td>2.47–4.60</td>
</tr>
<tr>
<td>5</td>
<td>&gt; 4.60</td>
</tr>
</tbody>
</table>

The boundaries for small $h$ have greater resolution, since the properties of the ice pack are especially sensitive to the amount of thin ice [Maykut, 1982].

Vertical heat conduction through each category is computed across four vertical layers of equal thickness, where internal temperatures are represented at the center, and conductivities and energy fluxes at the layer interfaces. The top and bottom surface ice temperatures constitute the vertical temperature boundary conditions.

In the horizontal, the Arakawa B-grid spatial discretization is used for staggering the variables (Figure 5). The thermodynamic state variables are at the center of the grid cell, and the velocity is defined at the grid cell corner. The stress tensor, rates of strain, and viscosities, defined bilinearly across the grid cell, are calculated from the values at the corners.

The time step, $\Delta t$, is constrained by the stability of transport, both horizontally and in thickness space. The horizontal transport time step bound under conservative remapping is given by

$$\Delta t < \frac{\min(\Delta x, \Delta y)}{2 \max(u, v)}.$$ (74)

The bound on the time step in terms of transport in thickness space is relatively less restricted:

$$\Delta t < \frac{\min(\Delta H)}{2 \max f},$$ (75)

where $\Delta H$ is the difference between category boundaries (Table 3) and $f$ is the rate of thermodynamic growth.

The thermodynamics component is stable regardless of the time step. The stability of the dynamics component does not depend on $\Delta t$, but on $\Delta t_e$ sufficiently resolving the damping time-scale, $T_{ew}$, given by Equation (67). The ratio

$$\Delta t_e : T_{ew} : \Delta t = 1 : 40 : 120$$ (76)
FIGURE 5. Variable staggering on the Arakawa B-grid. $\Delta x$ and $\Delta y$ denote grid spacing in $x$- and $y$-directions, respectively. Thermodynamic state variables sea ice area, $a_n$, sea ice volume, $v_{nl}$, sea ice internal energy, $e_{nl}$, snow volume, $v_{sn}$, snow/ice surface temperature, $T_{sfn}$, and compactness, $\sigma$, are at the center of the grid cell. Sea ice velocity, $u_{i,j}$, is defined at the north-east corner.

adequately provides stability and efficiency [Hunke and Libscomb, 2004]. For this study $E_o = 0.36$ and the dynamics component is subcycled 120 times per time step.

II.2 THE OCEAN MODEL

The ocean model used in this study is the Regional Ocean Modeling System (ROMS) version 2.1, the computational nonlinear kernel of which is described in Shchepetkin and McWilliams [2003, 2005]. ROMS, based on the Rutgers University S-Coordinate Primitive Equation Ocean Circulation Model (SCRUM) [Song and Haidvogel, 1994; Hedström, 1997], is a split-explicit time-stepping, free surface, hydrostatic, $s$ (terrain-following) vertical coordinate, primitive equation ocean model, numerically improved for computational efficiency in distributed and shared memory parallel platforms.

ROMS has been used for applications from the basin to coastal scales [e.g., Haidvogel et al., 2000; Marchesiello et al., 2001; Dinniman et al., 2003]. The model features several options for horizontal advection schemes [Shchepetkin and McWilliams, 2003,}
2005] and subgrid-scale parameterizations of lateral [Haidvogel and Beckmann, 1999] and vertical [Warner et al., 2005] mixing. The atmosphere-ocean interaction is based on the bulk parameterization adapted from the Coupled Ocean-Atmosphere Response Experiment (COARE) v2.0 algorithm [Fairall et al., 1996]. ROMS also includes Lagrangian particle tracking and coupled models for biogeochemical, bio-optical, and sediment applications.

The ocean model formulation is described in Section II.2.1 and Section II.2.2 provides the details of the numerical implementation.

II.2.1 Model Formulation

The free surface primitive equations included in ROMS are solved using potential temperature, salinity, and the oceanic equation of state, over variable topography and coastline using stretched terrain-following (s) vertical coordinates. The vertical coordinate permits enhanced resolution near the surface and the bottom, and the orthogonal curvilinear horizontal coordinates conform to irregular lateral boundaries.

The hydrostatic primitive equations for total and vertically integrated momentum are individually integrated using a split-explicit time-stepping for computational efficiency and then the barotropic (fast) and baroclinic (slow) modes are coupled. Both two- and three-dimensional equations are time-discretized by a third-order accurate predictor (Adams-Bashfor) and corrector (Adams-Molton) time-stepping scheme which allows larger time steps with its enhanced stability. The third-order, upstream biased horizontal advection scheme is dissipative with velocity-dependent hyperdiffusion, therefore requiring a small background explicit horizontal momentum mixing [Shchepetkin and McWilliams, 1998]. Laplacian (harmonic) mixing of momentum and tracers are along geopotential surfaces. The vertical mixing of momentum and tracers is parameterized by a nonlocal closure scheme based on the K-profile parameterization (KPP) by Large et al. [1994] expanded to include the surface and bottom boundary layers. Open boundary conditions are handled by a two-dimensional radiation scheme with adaptive nudging [Marchesiello et al., 2001].

Equations of motion

The primitive equations in the Cartesian coordinate system are given by
Table 4. Definition and values of the parameters and physical constants used in the ocean model

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \rho_0 )</td>
<td>Mean density of seawater</td>
<td>1025 kg m(^{-3})</td>
</tr>
<tr>
<td>( g )</td>
<td>Acceleration of gravity</td>
<td>9.81 m s(^{-2})</td>
</tr>
<tr>
<td>( c_w^a )</td>
<td>Specific heat of seawater</td>
<td>4218 J kg(^{-1}) K(^{-1})</td>
</tr>
<tr>
<td>( \gamma_1 )</td>
<td>Linear bottom drag coefficient</td>
<td>0 m s(^{-1})</td>
</tr>
<tr>
<td>( \gamma_2 )</td>
<td>Quadratic bottom drag coefficient</td>
<td>3 \times 10(^{-3})</td>
</tr>
<tr>
<td>( \nu_m^2 )</td>
<td>Harmonic horizontal viscosity</td>
<td>0.1 m(^2) s(^{-1})</td>
</tr>
<tr>
<td>( \nu_{\Theta}^2, \nu_S^2 )</td>
<td>Harmonic horizontal diffusivity of tracers</td>
<td>5 m(^2) s(^{-1})</td>
</tr>
<tr>
<td>( c_\rho )</td>
<td>Linear dependence of seawater density to depth</td>
<td>4.78 \times 10(^{-3}) kg m(^{-4})</td>
</tr>
<tr>
<td>( \gamma_T )</td>
<td>Turbulent exchange coefficient for heat</td>
<td>10(^{-4}) m s(^{-1})</td>
</tr>
<tr>
<td>( \gamma_S )</td>
<td>Turbulent exchange coefficient for salt</td>
<td>5.05 \times 10(^{-7}) m s(^{-1})</td>
</tr>
</tbody>
</table>

\(^a\) The specific heat of seawater, \( c_w \), is modified from the default value (3985 J kg\(^{-1}\) K\(^{-1}\)) to match the one used in the sea ice model (see Table 2).

\[
\frac{\partial u}{\partial t} + u \cdot \nabla u - f v = -\frac{\partial \phi}{\partial x} + F_u + D_u, \quad (77)
\]

\[
\frac{\partial v}{\partial t} + u \cdot \nabla v + f u = -\frac{\partial \phi}{\partial y} + F_v + D_v, \quad (78)
\]

\[
0 = -\frac{\partial \phi}{\partial z} - \frac{\rho_w g}{\rho_0}, \quad (79)
\]

\[
\frac{\partial \Theta}{\partial t} + u \cdot \nabla \Theta = F_\Theta + D_\Theta, \quad (81)
\]

\[
\frac{\partial S}{\partial t} + u \cdot \nabla S = F_S + D_S, \quad (82)
\]

\[
\rho_w = \rho_w(\Theta, S, P), \quad (83)
\]

where \( \nabla = (\partial/\partial x, \partial/\partial y, \partial/\partial z) \), \( f = f(y) \) is the Coriolis parameter, \( \rho_0 \) is the mean density used in the Boussinesq approximation (Table 4), \( g \) is the acceleration of gravity (Table 4), \( P = -\rho_0 g z \) is the total pressure, and \( \phi = \phi(x, y, z, t) = P/\rho_0 \) is the dynamic pressure. The terms denoted by \( F \) and \( D \) are the forcing and diffusive terms, respectively, for the model prognostic variables velocity, \( u_w = (u, v, w) \), potential temperature, \( \Theta(x, y, z, t) \), and salinity, \( S(x, y, z, t) \). The equation of state, given by (83), follows Jackett and Mcdougall [1995] to calculate the in situ density \( \rho_w(x, y, z, t) \).

Vertical boundary conditions at the surface \( z = \eta(x, y, t) \), where \( \eta \) is the sea surface
elevation, are given by

\[ \nu_m \frac{\partial u}{\partial z} = \tau^x_a; \quad \nu_m \frac{\partial v}{\partial z} = \tau^y_a; \quad \nu_\theta \frac{\partial \theta}{\partial z} = \frac{Q_{top}}{\rho_0 c_w}; \quad \nu_s \frac{\partial S}{\partial z} = \frac{(E - P)}{\rho_0} S, \quad (84) \]

\[ w = \frac{\partial \eta}{\partial t}, \quad (85) \]

where \( \nu_m \) is the vertical viscosity, \( \nu_\theta \) is the vertical diffusivity of potential temperature, \( \nu_s \) is the vertical diffusivity of salt, \( \tau^x_a \) and \( \tau^y_a \) are the surface wind stress components in the \( x \)- and \( y \)-directions, respectively, \( Q_{top} \) is the surface heat flux, \( E \) is the evaporation, \( P \) is the precipitation, and \( c_w \) is the specific heat of seawater (Table 4). Vertical boundary conditions at the bottom, \( z = -h(x, y) \), where \( h \) is the bottom depth, are given by

\[ \nu_m \frac{\partial u}{\partial z} = \tau_b^x; \quad \nu_m \frac{\partial v}{\partial z} = \tau_b^y; \quad \nu_\theta \frac{\partial \theta}{\partial z} = 0; \quad \nu_s \frac{\partial S}{\partial z} = 0, \quad (86) \]

\[ w = u_w \cdot \nabla h, \quad (87) \]

where \( \tau_b^x \) and \( \tau_b^y \) are the bottom stress in the \( x \)- and \( y \)-directions, respectively. The prescribed bottom stress components are given by

\[ \tau_b^x = (\gamma_1 + \gamma_2 \sqrt{u^2 + v^2}) u, \quad (88) \]

\[ \tau_b^y = (\gamma_1 + \gamma_2 \sqrt{u^2 + v^2}) v, \quad (89) \]

where \( \gamma_1 \) and \( \gamma_2 \) are the linear and quadratic bottom stress coefficients, respectively (Table 4). Quadratic bottom drag formulation is chosen by setting the linear drag coefficient, \( \gamma_1 \) to zero.

**Terrain-following (s) vertical and horizontal curvilinear coordinates**

The \( s \) stretched vertical coordinate system is a transformation of the vertical coordinate, \( z \), to \( s = s(x, y, z) \), \( -1 \leq s \leq 0 \), which essentially removes the variability of the bathymetry, \( z = -h(x, y) \). The vertical coordinate of the numerical model then fits smoothly to the irregular shape of the bathymetry, avoiding the false effects of the discontinuous, stepwise representation of the bottom surface and the side-walls [Haidvogel and Beckmann, 1999]. Following Song and Haidvogel [1994], \( z \) is stretched nonlinearly to have higher resolution near the surface and the bottom to better represent the mixed layer, the thermocline, and the bottom boundary layer. The transformation used is

\[ z = \eta(1 + s) + h_c s + (h - h_c) C(s), \quad -1 \leq s \leq 0, \quad (90) \]
where \( h_c \) is the smaller of the depth above which higher resolution is desired and the minimum depth, and \( C(s) \) is the stretching curve, defined as

\[
C(s) = (1 - b) \frac{\sinh(\theta s)}{\sinh \theta} + b \frac{\tanh(\frac{1}{2} s + \frac{1}{2}) - \tanh(\frac{1}{2} \theta)}{2 \tanh(\frac{1}{2} \theta)},
\]  

(91)

where \( \theta \), \( 0 < \theta \leq 20 \), and \( b \), \( 0 \leq b \leq 1 \), are the surface and bottom control parameters, respectively.

Defining \( H_z \equiv \frac{\partial z}{\partial s} \), the vertical velocity, \( \Omega \), in \( s \) coordinates is given by

\[
\Omega(x, y, s, t) = \frac{1}{H_z} \left[ w - (1 + s) \frac{\partial \eta}{\partial t} - u \frac{\partial z}{\partial x} - v \frac{\partial z}{\partial y} \right],
\]  

(92)

where

\[
w = \frac{\partial z}{\partial t} + u \frac{\partial z}{\partial x} + v \frac{\partial z}{\partial y} + \Omega H_z.
\]  

(93)

Redefining the ocean velocity and the divergence operator as

\[
\mathbf{u}_w = (u, v, \Omega),
\]  

(94)

\[
\mathbf{u}_w \cdot \nabla = u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} + \Omega \frac{\partial}{\partial s},
\]  

(95)

the primitive equations (77)-(83) are transformed into the dynamic equations

\[
\frac{\partial u}{\partial t} + \mathbf{u}_w \cdot \nabla u - f_v = - \frac{\partial \phi}{\partial x} - \frac{g \rho}{\rho_0} \frac{\partial z}{\partial x} - g \frac{\partial \eta}{\partial x} + F_u + D_u,
\]  

(96)

\[
\frac{\partial v}{\partial t} + \mathbf{u}_w \cdot \nabla v + f_u = - \frac{\partial \phi}{\partial y} - \frac{g \rho}{\rho_0} \frac{\partial z}{\partial y} - g \frac{\partial \eta}{\partial y} + F_v + D_v,
\]  

(97)

\[
0 = - \frac{\partial \phi}{\partial s} - \frac{\rho_w g H_z}{\rho_0},
\]  

(98)

\[
\frac{\partial H_z}{\partial t} + \nabla \cdot (H_z \mathbf{u}_w) = 0.
\]  

(99)

\[
\frac{\partial \Theta}{\partial t} + \mathbf{u}_w \cdot \nabla \Theta = F_\Theta + D_\Theta,
\]  

(100)

\[
\frac{\partial S}{\partial t} + \mathbf{u}_w \cdot \nabla S = F_S + D_S,
\]  

(101)

\[
\rho_w = \rho_w(\Theta, S, P),
\]  

(102)

resulting in \( \Omega = 0 \) as boundary conditions on vertical velocity at the surface \( (s = 0) \) and the bottom \( (s = -1) \).

The terrain-following coordinate system is sensitive to topography and pressure.
gradient errors arise for large $|\nabla h|$ due to the splitting of the pressure gradient term into an along-$s$ component and a "hydrostatic correction" [Haidvogel and Beckmann, 1999; Hedström, 2000]. The form of the pressure gradient (PG) that accounts for horizontal differences taken along the $s$ coordinate is given by

$$\text{PG} = H_z \nabla \psi + \left[ \frac{\rho_w g H_z}{\rho_0} \nabla z + g H_z \nabla \eta \right].$$

(103)

The hydrostatic correction terms in brackets are used to remove the pressure variations due to hydrostatic pressure changes along $s$-surfaces, however, they tend to cancel [Haidvogel and Beckmann, 1999; Shchepetkin and McWilliams, 2003]. In addition to topography smoothing that requires $\Delta h/2h \leq 0.2$ for reliable results, the numerical algorithm used to calculate the pressure gradient follows the conservative parabolic-spline vertical discretization method of Shchepetkin and McWilliams [2003] which reduces the baroclinic pressure gradient truncation error.

In the horizontal, the dynamic equations (96)-(102) are evaluated using orthogonal curvilinear coordinates $\mathcal{X}(x, y)$ and $\mathcal{Y}(x, y)$, where the relationship between the differential distances $(\Delta \mathcal{X}, \Delta \mathcal{Y})$ for each grid cell and the physical arc length $(\Delta S)$ is given by

$$\begin{align*}
(dS)_{\mathcal{X}} &= \left( \frac{1}{m} \right) d\mathcal{X}, \\
(dS)_{\mathcal{Y}} &= \left( \frac{1}{n} \right) d\mathcal{Y},
\end{align*}$$

(104)

where $m(\mathcal{X}, \mathcal{Y})$ and $n(\mathcal{X}, \mathcal{Y})$ are the scale factors (i.e. metrics). Constant metrics result in Cartesian coordinates, whereas variable metrics are used to formulate spherical coordinates [Haidvogel and Beckmann, 1999].

**Vertical and horizontal mixing**

The KPP vertical mixing scheme [Large et al., 1994] is a first order scheme that distinguishes between the surface boundary layer and the ocean interior regimes. Above a calculated oceanic surface boundary layer depth, $h_{abl}$, the formulation is based on boundary layer similarity theory. In the interior, the parameterization includes the effects of local Richardson number instability due to resolved vertical shear, double diffusion, and internal wave breaking.
In the surface boundary layer, vertical viscosity and diffusivities are expressed as

\[ \nu_\psi = h_{ssl} w_\psi(\sigma) G_\psi(\sigma), \]  

(106)

where the subscript \( \psi \) stands for momentum \( m \), potential temperature \( \Theta \), or salinity \( S \), \( \sigma \) is the non-dimensional coordinate between 0 and 1, \( w_\psi \) is a turbulent velocity scale, and \( G_\psi \) is a non-dimensional shape function. Oceanic surface boundary layer depth, \( h_{ssl} \) is the minimum of the Ekman depth, the Monin-Obukhov depth, and the shallowest depth that the bulk Richardson number, \( Ri_c \), reaches its critical value of 0.3. The Ekman depth is given by \( h_e = R_{i0} u_*/f \), where \( R_{i0} = 0.7 \) is the critical gradient Richardson number below which turbulent mixing occurs and \( u_* = \sqrt{(\tau_x)^2 + (\tau_y)^2}/\rho_0 \) is the friction velocity. The Monin-Obukhov depth is given by \( L = u_*^2/(\kappa B_f) \), where \( \kappa = 0.4 \) is the von Kármán constant and \( B_f \) is the buoyancy flux.

In the interior, the profile of overall effective viscosity and diffusivities, \( \nu_\psi(z) \), become

\[ \nu_\psi(z) = \nu^{s}_\psi(z) + \nu^{d}_\psi(z) + \nu^{w}_\psi(z), \]  

(107)

where the superscripts \( s \), \( d \), and \( w \) denote shear mixing, double diffusive mixing, and internal wave generated mixing, respectively. The shear mixing term is given by

\[ \nu^s_\psi(z) = \begin{cases} 
\nu_0 & Ri_g < 0, \\
\nu_0 \left[ 1 - \left( \frac{Ri_g}{Ri_0} \right)^2 \right]^3 & 0 < Ri_g < Ri_0, \\
0 & Ri_g > Ri_0,
\end{cases} \]  

(108)

where \( \nu_0 = 5.0 \times 10^{-3} \text{ m}^2\text{s}^{-1} \) and \( Ri_g \) is the gradient Richardson number. which is given by

\[ \frac{\partial b}{\partial z} \left( \frac{\partial b}{\partial z} \right)^2 + \left( \frac{\partial b}{\partial z} \right)^2, \]  

(109)

where \( b = g(1 - \rho_w/\rho_0) \) is buoyancy. The double diffusive mixing term, parameterized in general from laboratory and field data, is calculated differently for the salt fingering and diffusive convection cases [Large et al., 1994]. Internal wave generated viscosity and diffusivities are the constants \( 1.0 \times 10^{-4} \text{ m}^2\text{s}^{-1} \) and \( 1.0 \times 10^{-5} \text{ m}^2\text{s}^{-1} \), respectively [Large et al., 1994].

The Laplacian of a scalar \( \xi \), i.e. any of \( u, v, \Theta, \) and \( S \), in curvilinear coordinates is given by

\[ \nabla^2 \xi = mn \left[ \frac{\partial}{\partial X} \left( \frac{m}{n} \frac{\partial \xi}{\partial X} \right) + \frac{\partial}{\partial Y} \left( \frac{n}{m} \frac{\partial \xi}{\partial Y} \right) \right], \]  

(110)
Horizontal harmonic mixing is computed by multiplying (110) by $\frac{\nu_2^{\psi}H_z}{mn}$, where $\nu_2^{\psi}$ is the harmonic horizontal viscosity or diffusivity (Table 4).

**Ice shelf processes**

The dynamic and thermodynamic coupling between the ice shelf and the underlying ocean follows Dinniman et al. [2007]. The ice shelf is assumed to be in isostatic equilibrium and the hydrostatic pressure ($P_h$) at the base of the ice shelf is calculated by the integral from the mean sea level to the base of the ice shelf of the density of the seawater replaced by ice:

$$P_h = g\rho_w(N_w - 0.5c_pH_i)H_i,$$

where $\rho_w(N_w)$ is the density of the ocean surface layer, $c_p$ is the constant linear dependence of seawater density to depth (Table 4), and $H_i$ is the ice shelf draft. The stress terms in the $x$- and $y$-directions between the ice shelf and the seawater are given by the same formulation as the bottom stress components (Equations (88) and (89)), and are applied to the top three ocean levels as a body force.

The heat flux into the ocean is given by

$$\rho_i(L_F - c_0\Delta T)\frac{\partial h}{\partial t} = \rho_{mld}c_w\gamma_T(T_b - \Theta_{N_w}),$$

where $\Delta T$ is the temperature difference between the top (minimum of -1.95°C and air temperature) and bottom (-1.95°C) surfaces of the ice shelf, $\frac{\partial h}{\partial t}$ is the rate of melting or freezing at the ice shelf base, expressed as the equivalent change in ice thickness, $\rho_{mld}$ is the density of seawater in the mixed layer, $\gamma_T$ is the turbulent exchange coefficient for heat (Table 4), $T_b$ is the temperature at the ice shelf base, and $\Theta_{N_w}$ is the potential temperature of the ocean surface layer. The ice shelf base temperature ($T_b$) is given by

$$T_b = 0.0939 - 0.057S_b + 7.6410 \times 10^{-4}z,$$

where $S_b$ is the salinity at the ice shelf base calculated from the conservation of salt at the ice shelf-ocean boundary given by

$$\rho_iS_b\frac{\partial h}{\partial t} = \rho_{mld}\gamma_S(S_b - S_{N_w}),$$

where $\gamma_S$ is the turbulent exchange coefficient for salt (Table 4) and $S_{N_w}$ is the salinity
Figure 6. Variable staggering on the Arakawa C-grid. $\Delta x$ and $\Delta y$ denote grid spacing in $x$- and $y$-directions, respectively. The equation of state is evaluated at the center of the grid cell to compute water density, $\rho_w$. The surface elevation, $\eta$, the Coriolis term, $f$, and the vertical velocity in $s$ coordinates, $\Omega$, are also evaluated at the grid cell center, whereas the horizontal velocity components, $u_{i,j}$ and $v_{i,j}$, are computed at the midpoints of the west and south edges, respectively.

II.2.2 Numerical Solution

In the vertical, the water column is discretized into layers, numbered 1 to $N_w$ from the bottom to the surface. The number of layers, $N_w$, is uniform for each horizontal grid point. For each layer, $k$, the $s$ coordinate at the top surface of the layer is $s_k = (k - N_w)/N_w$.

The default horizontal and vertical discretizations use centered, second-order finite difference approximations. The variables are horizontally arranged according to the Arakawa C-grid (Figure 6), a staggering well suited for horizontal resolution ($\Delta x'$, $\Delta y'$) fine compared to the first radius of deformation [Arakawa and Lamb, 1977]. The vertical grid is also staggered: vertical velocity in the $z$ coordinate is evaluated at $s_k$, whereas the variables shown in Figure 6 are evaluated at the "midpoint" $s$ coordinate, i.e. at $s_m = s_k - 1/(2N_w)$. An optional higher order vertical stencil via a conservative, parabolic spline reconstruction of the vertical derivatives is also
implemented \cite{Shchepetkin and McWilliams, 2005}.

The time step $\Delta t$ of the ocean model is determined based on the stability of the internal processes. Waves that propagate at a speed of $\sqrt{gh}$ due to the presence of free surface imposes a more stringent time step limit on the barotropic mode. Therefore, the barotropic time step is subcycled within each $\Delta t$ to evolve the free surface and depth-integrated momentum equations. For this study, the number of subcycling time steps is 30. The barotropic fields are time-averaged by a cosine-shape time filter, centered at the new time level, and the average values replace those obtained with the baroclinic time step. The two-way time-averaging procedure satisfies continuity, thereby guaranteeing exact volume conservation and constancy preservation of tracers \cite{Shchepetkin and McWilliams, 2005}. This feature allows the model to remain numerically stable and accurate when the sea level change is a significant fraction of the unperturbed depth. The time-stepping employed in the ocean model allows an increase in the permissible time step size, which offsets the cost of the predictor-corrector algorithm. This scheme, with a dissipation-dominant truncation error, also closely couples the momentum and tracer fields and suppresses the computational modes.

II.3 COUPLED MODEL IMPLEMENTATION

The sea ice-ocean model dynamically couples the respective models via a flux coupler that exchanges momentum, heat, and buoyancy (salt and freshwater) fluxes between the models at any given coupling time interval (Section II.3.1). The coupled model is implemented both as a one-dimensional time-dependent ($z$-t) model (Section II.3.2) and a three-dimensional model of the Ross Sea, Antarctica (Section II.3.3).

II.3.1 Sea Ice-Ocean Coupling

The sea ice model uses the atmospheric input (Figure 7) for the calculations presented in Section II.1. The open water flux exchange between the atmosphere and the ocean follows the COARE bulk parameterization of air-sea fluxes \cite{Fairall et al., 1996}. Shortwave radiation is calculated by the ocean model \cite{Dinniman et al., 2003} and is passed as a forcing term to the sea ice model via the flux coupler (Figure 7).

When the seawater temperature drops below its freezing point, frazil ice forms in the water column and floats to the surface. The ocean model calculates the freez-
Figure 7. Schematic of model atmospheric forcing and sea ice-ocean coupling fluxes. Atmospheric forcing is same for both the ocean and sea ice models, except surface wind stress is used by the ocean model only. Fluxes in the "OCEAN MODEL" and "SEA ICE MODEL" boxes are calculated by the respective model component and exchanged via the flux coupler at the coupling time step.

The freezing/melting potential of the water column as

$$F_{frzmlt} = \left[ \rho_w c_w \sum_{n_w}^{N_w} (T_f - \Theta_{n_w}) \Delta h_{n_w} \right] / \Delta t, \quad (115)$$

where $n_w$ denotes the ocean depth level index, $\Delta h_{n_w}$ is the thickness of the layer, and $\Theta_{n_w}$ is the potential temperature of the layer. The freezing point calculation for the seawater takes into account the pressure effect on freezing, following Steele et al. [1989], as

$$T_f = T_f(S, z) = \mu S + 0.000759 z. \quad (116)$$

If $F_{frzmlt} > 0$ then the ocean model assumes the amount of frazil ice implied by the freezing potential is formed. The temperature of each oceanic layer that forms frazil
ice is restored to its freezing point. Brine associated with freezing, given by

\[ \Delta S = \frac{(S_{nw} - S_l) \Delta t F_{frzmlt}}{\rho_0 \Delta h_{nw} L_i}, \]

where \( S_l = 4 \) is the constant reference salinity of sea ice, is added to the salinity of that layer.

The sea ice model receives \( F_{frzmlt} \) as input and, for positive \( F_{frzmlt} \), new ice is formed in open water. Frazil ice in excess of the amount that would fit in the thinnest ice category in the open water is distributed evenly over the entire grid cell. Negative \( F_{frzmlt} \) is used for basal and lateral melting. The portion of the melting potential used by the sea ice model is returned to the ocean via the coupler to let the ocean model adjust its heat budget. The freshwater and salt fluxes associated with congelation ice formation and ice melting (top, basal, and lateral) are also returned to the ocean as buoyancy fluxes.

### The coupling interface

The coupling of models is implemented using the Weather Research and Forecasting Input/Output Application Programming Interface Model Coupling Toolkit (WRF I/O API MCT), a package-independent I/O API, which is a part of the WRF Software Architecture [Michalakes et al., 2004]. WRF I/O API is implemented using the MCT v2.0.1 of the Argonne National Laboratory (ANL) [Larson et al., 2005; Jacob et al., 2005]; the Message Passing Environment Utilities (MPEU) v1.5 distributed as part of the MCT; Nansen Environmental and Remote Sensing Center (NERSC) Multi Program-Component Handshaking Library (MPH) v4 [He and Ding, 2005]; and Network Common Data Form (NetCDF) I/O library of Unidata/University Corporation for Atmospheric Research (UCAR) [Rew and Davis, 1990].

Coupling between the ocean and sea ice models is loose, i.e. models may have different grids and time steps, and coupling can be done at any time interval. If the models have different grids, conservative remapping and interpolation of area-averaged fields between grids are computed once offline by Spherical Coordinate Remapping and Interpolation (SCRIP) package of LANL [Jones, 1999] and are provided as input to the coupler at the initialization stage. Following the exchange of fluxes, the ocean model interpolates the momentum fluxes from B-grid to C-grid, and the sea ice model vice versa. The ocean model weighs the fluxes by ice concentration in the bulk flux calculations.
**Coupled model time-stepping**

Parallel computer processors allocated for the model runs are split between the sea ice and ocean models. Models run simultaneously and coupling is concurrent (Figure 8). The coupled sea ice-ocean model code follows the Earth System Modeling Framework (ESMF) conventions for model coupling with each model carrying out the integration in *initialize, run, and finalize* stages [ESMF Joint Specification Team, 2006].

The time interval at which forcing and coupling fluxes are sent to and received from the coupling interface is termed the “coupling time step”. Upon receiving the atmospheric forcing fields, the vertical thermodynamics calculation updates the sea ice surface states and sea ice-atmosphere fluxes that are required if an atmospheric model is also coupled. The coupling time step is split into two intervals for computational efficiency, i.e. once calculated by the vertical thermodynamics module, the sea ice states and atmosphere-ice ocean fluxes are returned to the coupler so that the ocean and the atmosphere models can run in parallel during the remainder of their respective time steps. Sea ice thickness distribution and sea ice-ocean fluxes are computed following the exchange with the coupler. Although this study does not include an atmospheric model, the coupling mechanism is not altered to preserve computational efficiency.

II.3.2 Vertical Time-dependent (z-t) Model

In order to isolate and analyze the model sensitivity to the vertical processes, a one-dimensional version of the coupled model is implemented. The vertical time-dependent model is configured by reducing the horizontal grid in the x- and y-directions to the minimum number of grid points required by the numerical stencils and applying doubly-periodic boundary conditions to remove horizontal gradients. Although the z-t model includes sea ice dynamics that drive sea ice advection and mechanical redistribution, double periodicity removes sea ice divergence and convergence, rendering the model essentially thermodynamic.

In the absence of horizontal gradients, lateral heat and salt exchange at depth is represented by nudging the temperature and salinity toward the initial conditions or a prescribed profile. Nudging is implemented in increasing strength with depth. No nudging is applied to the surface layer above the pycnocline to let the model determine the mixed layer structure, while sub- pycnocline simulated temperature and salinity values are replaced by the initial conditions at each time step. Hence, the new values \((\Theta_{\text{new}}, S_{\text{new}})\) computed from the initial values \((\Theta_{\text{init}}, S_{\text{init}})\) and model calculated
Figure 8. Coupled model flowchart. “MASTER” denotes the master program that spawns the ocean and sea ice components, denoted by “OCEAN” and “SEA ICE”, respectively. The arrows denote the sequence of integration of respective model processes. “BC” denotes “boundary conditions”. “ITD” stands for “ice thickness distribution”. “3D” denotes calculations in three spatial dimensions. “RHS” is the right-hand-side of the equations of motion and of tracer conversation.
Figure 9. The nondimensional nudging coefficient \( \nu_{\text{ndg}}(z) \) for \( z_c = 250 \text{ m} \) and \( D_c = 20 \text{ m} \).

values \( (\Theta_{\text{sim}}, S_{\text{sim}}) \) using the depth-dependent nudging coefficient, \( \nu_{\text{ndg}} = \nu_{\text{ndg}}(z) \), are given by

\[
\begin{align*}
\Theta_{\text{new}}(z) & = \Theta_{\text{sim}}(z) + \nu_{\text{ndg}}[\Theta_{\text{init}}(z) - \Theta_{\text{sim}}(z)], \\
S_{\text{new}}(z) & = S_{\text{sim}}(z) + \nu_{\text{ndg}}[S_{\text{init}}(z) - S_{\text{sim}}(z)].
\end{align*}
\]  

(118)  

(119)

The nudging coefficient, \( \nu_{\text{ndg}} \), increases from 0 to 1 through the pycnocline and is calculated as

\[
\nu_{\text{ndg}}(z) = \frac{1}{2} \left[ \tanh \left( \frac{z - z_c}{D_c} \right) + 1 \right],
\]

(120)

where \( z_c \) is the midpoint depth and \( D_c \) is a coefficient determining the thickness of the transition layer. The vertical profile of the nudging coefficient for \( z_c = 250 \text{ m} \) and \( D_c = 20 \text{ m} \) applied in the reference simulation of the z-t model is shown in Figure 9.

The z-t model, uses three interior grid points in \( (x, y) \) with grid spacing \( \Delta x = \Delta y = 20 \text{ km} \). Sea ice-ocean coupling is done at every physical time step, \( \Delta t = 1800 \text{ s} \).
II.3.3 Ross Sea Model

The Ross Sea model domain (Figure 10) extends past the continental shelf break northward and includes the cavity beneath the Ross Ice Shelf (RIS). The northern boundary of the domain is to the south of winter maximum ice extent. The water column thickness underneath the ice shelf and the ice shelf draft are obtained from the BEDMAP gridded model [Lythe et al., 2001] which uses the ETOPO2 bathymetry by Smith and Sandwell [1997]. The bathymetry for the open ocean is derived from the ETOPO5 database [NGDC, 1988]. Grid spacing is 5 km in both directions and the water column is represented by 24 vertical levels, concentrated at the surface and bottom boundary layers (Figure 11), yielding the thickness of the top level at the surface varying between 2 and 33.6 m.

Initialization and forcing

The sea ice-ocean coupled simulations are initialized from the fields obtained from a six-year spin-up of the uncoupled ocean model with imposed sea ice cover [Dinniman et al., 2007]. The initial and open boundary fields of temperature and salinity for the spin-up are from the World Ocean Atlas 2001 (WOA01) [Conkright et al., 2002]. Open boundary barotropic velocity conditions are relaxed to the Ocean Circulation and Climate Advanced Modelling Project (OCCAM) [Webb, 1996; Webb et al., 1998] global model output. Air temperature, sea level air pressure, and relative humidity forcing fields are from the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA-40) monthly climatologies [Uppala et al., 2005]. Precipitation rate is from the Xie and Arkin [1997] climatology and cloud cover is from the ISCCP data [Rossow and Schiffer, 1999]. Daily values of wind velocity and stress for 2000 and 2001 are obtained from the blended product of data from NASA/Jet Propulsion Laboratory (JPL) SeaWinds Scatterometer aboard the QuikSCAT and National Centers for Environmental Prediction (NCEP) analysis [Milliff et al., 2004]. Sea ice cover is imposed from the SSM/I monthly climatology generated using the NASA Team algorithm [Cavalieri et al., 1990]. Monthly climatological snow thickness is from the Advanced Microwave Scanning Radiometer for the Earth Observing System (AMSR-E) data [Kawanishi et al., 2003].

Initial sea ice cover for the simulations is the September monthly climatological field [Cavalieri et al., 1990] at 0.5 m thickness per sea ice area. The temperature, salinity, and barotropic velocity open boundary conditions for the coupled model simulations are the same as in spin-up. The sea ice velocity boundary conditions are no-slip (i.e. $u_i = 0$) along the coastlines and "no gradient" at the open boundaries. Sea ice
Figure 10. The Ross Sea model domain. Ross Ice Shelf and the Drygalski Ice Tongue are delineated by the thick borders. The contours represent the water column thickness (bathymetry for open water). Filled, solid contours are at 100 m intervals up to 1000 m and the dotted contour interval is 250 m. The inset bottom left shows the location of the model domain on the Antarctic continent. Ice shelf draft is plotted in the figure bottom right. Line contours are drawn for 300 and 350 m ice thickness. Geographical locations indicated are: RIS-Ross Ice Shelf, RoI-Ross Island, MS-McMurdo Sound, DIT-Drygalski Ice Tongue, TNB-Terra Nova Bay, CA-Cape Adare, PB-Pennell Bank, CC-Cape Colbeck, and RsI-Roosevelt Island.
at the eastern open boundary was imposed from the SSM/I monthly climatological concentration [Cavalieri et al., 1990] with a prescribed thickness of 0.5 m per unit sea ice area. The atmospheric forcing for the simulations is daily, averaged from six-hourly data (Figure 12). Air temperature, sea level air pressure, relative humidity, and cloud fraction are from the ECMWF ERA-40 dataset. Precipitation rate is from the NCEP/NCAR 40-Year Reanalysis Project [Kalnay et al., 1996]. Wind forcing, argued to be the primary forcing for sea ice motion [Steele et al., 1997], is also obtained from the ECMWF ERA-40 dataset. For the reference simulation, however, data for two ERA-40 nodes near the Terra Nova Bay (at 75°S, 162.50°E and 165°E) are replaced by the average of measurements from three automatic weather stations (AWS) Jennica (74.70°S, 164.09°E), Eneide (74.70°S, 164.09°E), and Rita (74.73°S, 164.03°E) [Copyright Italian National Research Program in Antarctica. All rights reserved], before interpolating onto the model grid.

Figure 11. The vertical coordinates of the ρ-points along a model grid section near 170°W.
Figure 12. Area averaged atmospheric daily forcing (2000-01) for open ocean grid points. (a) Wind vectors aligned with model coordinates, (b) air temperature, (c) sea level air pressure, (d) relative humidity, (e) cloud fraction, and (f) precipitation rate.

The AWS data are considerably different from the ERA-40 data in terms of both magnitude and direction (Figure 13). The seasonal cycle present in the ERA-40 data is not present in the AWS data, and the latter captures the katabatic winds which are reported to drive the Terra Nova Bay polynya [Bromwich and Kurtz, 1984].

**Simulations and analysis methods**

The reference simulation is forced by the wind field that uses the AWS data for the
Figure 13. Terra Nova Bay daily wind forcing vectors (2000-2001) aligned with model coordinates. (a) ECMWF ERA-40 and (b) AWS data averaged from stations Jennica, Eneide, and Rita.

Terra Nova Bay (Figure 13b). Sensitivity study of the coupled model is carried out by using the ECMWF ERA-40 winds unmodified for the region (Figure 13a). This case is termed the “ECMWF” run. The Ross Sea simulations start on September 15, forced by the 2000 data for the remainder of the year as additional spin-up. Forcing for the following two years of simulation, output of which is used for analysis, is from 2000-01 data. Sea ice and baroclinic time step is $\Delta t = 300\, s$ and coupling is done at hourly intervals. Model output of averaged and snapshot fields is saved at 5-day intervals. Ten grid points each into the east and west sides of the domain are excluded from analysis to limit the spurious effects of manipulated bathymetry and relaxation to imposed boundary conditions.

Simulated sea ice concentration is compared to the fields derived from SSM/I passive microwave data using the NASA Team algorithm [Cavalieri et al., 1996] for validation purposes. This is the recently released, updated version of the dataset from which the climatological fields for ocean model spin-up, coupled model initial sea ice cover, and coupled model eastern open boundary sea ice conditions are obtained [Cavalieri et al., 1990].

The Ross Sea shows substantial regional variability in terms of ocean properties and sea ice dynamics. Regional analysis of the model results are carried out by partitioning the domain to western and eastern shelves, continental slope, and abyssal components (Figure 14a). Terra Nova Bay and Ross Sea polynya regions as well as the Victoria
Land shelf are analyzed for dense water formation due to sea ice-related processes. The “TNB-RSP” section (Figure 14b) is used to analyze salinity cross-sections. The yearly mean Θ-S distribution and the transport rate time series of deep water masses out of the RIS section (i.e. northward), the TNB, RSP, and PLY regions, and out of the SHB section (i.e. northward towards the shelf break; Figure 14b) are also reported in the following chapter.

Vertically integrated volume flux rate of water masses over the model domain are calculated at 100 km (20 grid points) resolution to analyze their distribution and pathways. Yearly mean net flux rate of a specific water mass for each 100×100 km region is graphically represented by an arrow, based at the center of the area, the magnitude and direction of which is determined by the relative magnitudes of northward and eastward flux out of the region. Time series of the transport rate of deep water masses are calculated for Modified Circumpolar Deep Water/Lower Circumpolar Deep Water (MCDW/LCDW), Modified Shelf Water/Antarctic Bottom Water (MSW/AABW), High Salinity Shelf Water (HSSW), and Low Salinity Shelf Water...
Figure 15. Deep water mass definitions used in the analysis. MCDW/LCDW - Modified Circumpolar Deep Water/Lower Circumpolar Deep Water, MSW/AABW - Modified Shelf Water/Antarctic Bottom Water, HSSW - High Salinity Shelf Water, and LSSW - Low Salinity Shelf Water. The lines that confine the MCDW/LCDW in the Θ-S space are approximations to the 28.0 and 28.27 neutral density contours used by Whitworth et al. [1998]. MSW/AABW definition follows Stover [2006].
CHAPTER III

RESULTS

The Ross Sea model results presented in this chapter correspond to two years of simulation forced with datasets from 2000-01. For the reference simulation (forced with ECMWF and AWS winds), model volume weighted average salinity below 200 and 300 m (Figure 16), including that of the seawater in the ice shelf cavity, shows the effects of the seasonal sea ice cycle, and the drift in salinity over the course of two years is deemed reasonably small (\(~ 0.001 \text{ PSU yr}^{-1}\)) for analysis. Analyses of the simulated annual and seasonal sea ice-ice shelf-ocean dynamics for the whole domain are described in Section III.1. Regional analysis of sea ice-ocean dynamics, polynya processes, and water mass properties as well as results from a diagnostic study using the vertical time-dependent sea ice-ocean model are presented in Section III.2. Section III.3 provides analysis of model sensitivity to Terra Nova Bay wind forcing.

![Figure 16. Model volume weighted average ocean and ice shelf cavity seawater salinity below 200 m (solid) and 300 m (dotted).](image)

III.1 LARGE SCALE SEA ICE-ICE SHELF-OCEAN DYNAMICS

III.1.1 General Characteristics of the Sea Ice Cover

Time series of domain-averaged simulated sea ice concentration (Figure 17) shows the characteristic seasonal behavior of the Ross Sea sea ice cover: freezing onset in February-March, rapid areal growth to maximum concentrations in four to six
Figure 17. Ross Sea domain averaged sea ice concentration, sea ice and snow thickness. Simulated (solid) and observed (dotted) mean sea ice concentration, simulated mean sea ice (dash dot) and snow (dash dots) thickness are shown.

weeks, winter equilibrium from April through November, melting onset by November-December, and areal decline to summer minimum in January-February [Jacobs and Comiso, 1989]. Main differences between the simulated and observed sea ice concentration are overestimated compactness of the modelled field in winter while lacking observed variability and longer simulated sea ice season due to slower summer melting in spite of faster winter freezing. At the beginning of the analyzed time series, domain averaged sea ice concentration is higher than the observed by about a factor of two. Winter freezing starts at the beginning of March in both years of simulation, delayed by about two weeks compared to the satellite data, however, reaches the winter maximum at about the same time. Once at the seasonal maximum, about 95%, simulated winter mean sea ice concentration is consistently compact and not variable as the SSM/I data, which fluctuate between 80 and 95%. The model captures well the low and high sea ice covers in summers of 2000 and 2001, respectively. The simulated values of summer sea ice concentration minima also reflect the respective difference in air temperature forcing (Figure 12). Simulated mean sea ice thickness, which shows winter variability unlike concentration, is always less than 1 m, within the range of reported averages [Gordon and Huber, 1990; Jeffries and Adolphs, 1997; Markus, 1999; Wadhams, 2000; Tin et al., 2004]. Maximum mean snow thickness on sea ice is less than 20 cm, which also is in accordance with the observations [Jeffries and Adolphs, 1997; Massom et al., 2001].

Simple correlation for the time series of domain averaged simulated sea ice concentration and SSM/I data yields $r = 0.96$ for year 2000, 0.92 for year 2001, and 0.94
Figure 18. Simulated and observed yearly mean sea ice concentration and sea ice drift. (a, c) Simulated and (b, d) observed for 2000 (top row) and 2001 (bottom row). Bathymetry contour shown is the 1000 m isobath. Vectors are drawn every 10 grid points.

for two years. The mean (root-mean-squared) deviation of domain averaged sea ice concentration from the SSM/I data for two years is -7.05% (±11.68%). Simulated sea ice cycle lags about 10 days behind that of the SSM/I data (11-day lagged $r = 0.99$ and the root-mean-squared deviation is ±8.65% for two years).

Yearly mean horizontal distribution of simulated sea ice concentration shows the reduced sea ice cover along Ross Ice Shelf, Ross Sea polynya area, Terra Nova Bay,
Figure 19. Difference between yearly mean simulated and observed sea ice concentration and sea ice area drift. (a) 2000 and (b) 2001. Bathymetry contour shown is the 1000 m isobath. Vectors are drawn every 10 grid points.

and the region around the Ross Island and McMurdo Sound for both years of the simulation (Figure 18a, c). The mean (root-mean-squared) deviations of simulated sea ice concentration from observations for all grid points are -7.06% (±10.20%) and -7.04% (±12.71%) for years 2000 and 2001, respectively. Within the continental shelf, denoted by the 1000 m isobath, simulated sea ice concentration compares fairly well with the SSM/I data (Figure 18b, d) with the exception of the regions to the east of Terra Nova Bay and north of the eastern part of the Ross Ice Shelf. Over the deeper part of the ocean, the model overestimates the sea ice concentration, whereas the values in the eastern part of the domain are underestimated. Simulated yearly mean sea ice drift (Figure 18a, c) conforms to the general structure of the SSM/I-derived sea ice motion [Fowler, 2003] (Figure 18b, d) except at the open northern boundary where the SSM/I-derived sea ice motion is uniformly northward and over the Ross Sea polynya area where the simulated velocities are about 5 cm/s higher.

A comparison between yearly mean simulated and observed sea ice concentration and motion shows that there is a regional pattern to where the simulated sea ice cover deviates from the satellite data (Figure 19). The negative difference in sea ice concentration over the eastern part of the domain is mainly due to the specification
of the eastern open boundary sea ice conditions, whereas the positive difference to the east of Terra Nova Bay and over the northern part of the domain is driven by the simulated convergent sea ice motion. The positive difference in the northern part of the domain is mainly wind driven. The difference in sea ice area drift is obtained from the vector difference of simulated and SSM/I-derived sea ice motion scaled by the simulated and observed sea ice concentrations, respectively. Simulated sea ice area flux differs from the satellite product by a north-northwest bias over the western Ross Sea and the western open boundary of the domain. Over the northeastern corner of the domain, however, the difference is reversed and has a south-southeast bias.

Simulated yearly mean sea ice thickness favorably reproduces the general characteristics of the Ross Sea ice pack (Figure 20). Thick sea ice accumulates along the Victoria Land coast, to the north of Cape Adare upon export, and along the continental slope (Figures 2, 3). The thickness of sea ice entering the region from the Amundsen sea is underestimated due to an imposed thickness scaled by the climatological SSM/I sea ice concentration at the boundary. The effects of sea ice motion on the mass distribution is well simulated. Regions of sea ice divergence, such as

Figure 20. Simulated yearly mean sea ice thickness and sea ice volume drift. (a) 2000 and (b) 2001. Simulated sea ice volume drift is normalized by the grid cell area. Sea ice thickness values greater than 1 m are represented by the same contour level. Bathymetry contour shown is the 1000 m isobath. Vectors are drawn every 10 grid points.
the Ross Sea and Terra Nova Bay polynyas, are indicated by sea ice less than 20 cm thick, whereas areas of convergence, such as the northern part of the Victoria Land coast, may accumulate sea ice thicker than 1 m (maximum thickness 3.4 m). In the northern part of the domain, sea ice is thicker in 2001 than 2000, due to accumulation in accordance with the wind forcing.

Sea ice motion is driven primarily by wind forcing, however, ocean surface flow becomes as effective for speeds on the order of 10 cm/s (Figure 21). Within the continental shelf, surface currents are not strong and coherent enough to augment or counteract the effect of wind forcing on sea ice. Over the continental slope and deep northeastern region of the domain surface flow contributes substantially to sea ice drift. Two main mechanisms of sea ice motion are apparent in the simulated sea ice drift: (i) sea ice formed in the western part of the Ross Sea continental shelf is exported off the shelf and out of the model domain from the northwestern corner, and (ii) sea ice entering into the region from the southeast boundary exits the domain from the north and northeast edges.

Thermodynamic volume tendencies over the RSP and TNB regions are substantially higher than the rest of the domain (Figure 22). Over most of the domain there is net thermodynamic production of sea ice in both simulation years. By the end of simulation year 2000, most of the thermodynamically produced sea ice (Figure 22a) is removed by the dynamics, except in parts of the eastern shelf and along the Victoria Land to the north of Terra Nova Bay where the sea ice drift is convergent, and in the northern part of the domain where the northward component of wind forcing is not strong enough (Figure 22b). For the simulation year 2001, the net dynamic volume tendency along the middle section of the domain and to the north and east of the Terra Nova Bay results from the drift of the residual sea ice from the previous simulation year (Figure 22d).

A balance of time-integrated net thermodynamic sea ice production, sea ice import-export at the domain boundaries, and sea ice volume shows that the Ross Sea is a region of net sea ice formation and export (Figure 23). Sea ice export rate is lower at the beginning of each freezing season due to sea ice cover forming over the continental shelf initially and expanding north later in the season (Figure 23a). Additionally, sea ice exported later in the season is thicker. Annually, about 2000 km$^3$ of sea ice is formed in the region and only 15-20% melts locally (Figure 23b). Sea ice drift, primarily driven by the wind, causes about 80-90% of the thermodynamically formed sea ice to exit the region and melt elsewhere. Of the net ~3200 km$^3$ sea ice thermodynamically produced within the two year simulation period, about 2800 km$^3$
Figure 21. Yearly mean simulated sea ice area drift, wind forcing, and ocean surface flow. Sea ice drift (top row), wind forcing (middle row), and ocean surface flow (bottom row) for 2000 (a, b, c) and 2001 (d, e, f). Bathymetry contour shown is the 1000 m isobath. Vectors are drawn every 10 grid points.
is exported and 400 km$^3$ remains in the domain.

### III.1.2 Mean Circulation and Ocean-Ice Shelf Interaction

Vertically averaged ocean flow is topographically controlled (Figure 24). The slope current follows the steep shelf break which abruptly increases from 500 m to more than 2000 m. At the end of the slope a well developed current traverses the domain diagonally from the eastern boundary to the northern. The flow enters into the ice shelf cavity in the western part of the Ross Sea and continues southward along the
western grounding line. In the eastern part of the cavity the flow is eastward and out of the cavity near Cape Colbeck. Flow out of the ice shelf cavity also takes place at several locations along the eastern part of the ice shelf edge. The overall flow pattern does not show variability between the two simulation years.

Yearly mean ice shelf melt rate (Figure 25) has a spatial pattern closely related to the yearly mean vertically averaged flow, with minimal variability between the two simulation years. High rates of melting occur along the ice shelf edge and the northwestern and southern grounding regions, consistent with the path of the oceanic flow into the cavity and the circulation within. Regions of net basal freezing, predominantly in the eastern half of the ice shelf, are characterized by relatively thicker ice shelf draft, shorter water column, and sluggish vertically averaged flow.

III.1.3 Seasonal Sea Ice Cycle

During the winter months the Ross Sea ice cover is characterized by lower concentrations and volume over the Ross Sea and Terra Nova Bay polynya areas, and along the Ross Ice Shelf. In July of the first year of simulation (2000), the simulated sea ice cover, except for a relatively narrow strip along the Ross Ice Shelf edge, is between 80 and 100% over the whole domain (Figure 26a). The simulated concentration is
Figure 24. Simulated yearly mean vertically integrated velocity for (a) 2000 and (b) 2001, and water column thickness. Isobaths deeper than 1000 m are not shown. Vectors are drawn every 6 grid points.

15-25% higher than the satellite data over most of the Ross Sea polynya area and the Terra Nova Bay, and over smaller regions near Cape Colbeck (Figure 26c). To the north of the shelf break, differences between the simulated and satellite values are in the -5 to 15% range. For the same period, however, simulated sea ice thickness over the Ross Sea polynya area and the Terra Nova Bay is less than 50 cm, about 10-20 cm near the coast and the ice shelf (Figure 26e). Simulated sea ice thickness over the Ross Sea continental shelf, except in the east between 160°W and 170°W and south of Cape Adare, is significantly less than the deeper parts of the domain. The overall pattern of simulated sea ice drift compares well with the satellite-derived field (Figure 26a, c), however, simulated magnitudes are larger over the Ross Sea polynya area and the Terra Nova Bay, and the directions differ over the Pennell Bank, regions near Cape Adare, and the northeast corner of the domain.
In August of the same year of simulation, the difference between the simulated and observed sea ice concentration decreases over parts of the Ross Sea polynya area and the Terra Nova Bay, while increasing north of the 1000 m isobath (Figure 26d). Simulated sea ice cover is more compact over the western half of the continental shelf compared to July (Figure 26b), while decreasing in thickness (Figure 26f). Simulated sea ice motion captures the observed change in the drift pattern over the eastern half of the domain, again with the exception of the northeastern corner of the domain (Figure 26b, d). Simulated sea ice moves faster with a western bias compared to the observations over the same regions, namely the western half of the domain, as in July.

In September, over the western part of the domain, simulated sea ice concentration begins to decrease (Figure 27a), while the difference between the simulated and observed cover does not change significantly (Figure 27c). The northern extent of the simulated thinner ice is well north of the continental slope near the Iselin Bank and in the southern half of the eastern boundary (Figure 27e). During this period, there is an influx of thicker ice from the Amundsen Sea into the domain, as imposed by the conditions in the northern part of the eastern open boundary. Simulated sea ice continues to thicken along convergence regions, i.e. on the northern part of the Victoria Land coast and the eastern part of the continental slope where ocean surface
Figure 26. Monthly mean (a, b) simulated sea ice concentration and drift, (c, d) percent difference between the simulated and the observed sea ice concentration and observed sea ice drift, and (e, f) simulated sea ice thickness in July (top row) and August (bottom row) 2000. Vectors are drawn every 10 grid points.

Simulated sea ice concentration continues to decrease into October over most of the Ross Sea continental shelf by 5-10% from September (Figure 27b). Difference between the simulated and observed sea ice cover (Figure 27d) and simulated sea ice thickness distribution (Figure 27f) are not significantly different from the previous month. The observed change in the direction of monthly average sea ice drift along the continental slope is well captured by the model (Figure 27b, d).

In July of the second year of simulation (2001), simulated sea ice cover is more compact over western Ross Sea compared to July, 2000 (Figure 28a). Such difference between two years of sea ice cover is partially reflected in the observations (Figure 28c) as the simulated values over northern part of the Ross Sea polynya area deviate less from the observations. Simulated sea ice cover in August of the same year is less compact than that of August, 2000 (Figure 28b) and compares very well to observed values (Figure 28d). This month is marked by slower observed sea ice drift and the
model reproduces this feature well. Sea ice thickness distribution in both months is similar to the results from the previous year of simulation (Figure 28e, f), however ca. 20 cm thicker in the northeast part of the domain.

The northern and eastern parts of the domain experience a thickening of the simulated sea ice in September and October of the second year of simulation, although, to the north of the Ross Ice Shelf edge within the continental shelf, simulated sea ice is thinner (Figure 29e, f). Over the southern part of the eastern boundary region, the thickening is due to an influx of thicker sea ice imposed on the boundary. Terra Nova Bay simulated sea ice thickness also is about 10-20 cm thicker during this period unlike the previous year of simulation. Two coastal polynya openings along the Ross Ice Shelf edge for both months (Figure 29a, b), which were not produced in the previous year of simulation, have more open water area than observed (Figure 29c, d). Regions that show difference in direction and magnitude between the simulated and satellite-derived sea ice motion fields on the western part of the continental shelf also yield relatively greater difference in the sea ice concentration distribution.

Overall, simulated Ross Sea winter sea ice cover over in July through October for
simulation years 2000 (Figures 26, 27) and 2001 (Figures 28, 29) compare fairly well to satellite data over most of the domain. Regions with differences in the 15-25% range are mostly localized to the polynya areas and to coastal regions near Cape Colbeck with spatial patterns matching very closely with the distribution of simulated thin sea ice. Simulated sea ice drift captures the high-frequency variability as reflected in monthly means and spatial pattern compares well with the SSM/I-derived sea ice motion. Regions where simulated and satellite-derived sea ice drift do not compare well in direction and/or magnitude coincide with regions where the difference in sea ice concentration is also larger.

Opening of the coastal polynyas occurs in November and December. In November of the first year of simulation (2000), the observed Ross Sea polynya is larger than the simulated and the lowest sea ice concentrations are concentrated over the Pennell Bank region (Figure 30a, c). The simulated polynya opening shows lower sea ice concentrations compared to the satellite data for the same area and the simulated sea ice drift is highly convergent to the north of the region. A similar effect, where the simulated and observed polynya sizes are different and the simulated sea ice drift is convergent, is seen around Cape Colbeck. Simulated sea ice thickness coincident
with the coastal polynyas along the Ross Ice Shelf edge, to the north of McMurdo Sound, and the Terra Nova Bay is less than 30 cm (Figure 30e).

The shape of simulated polynya openings in December (Figure 30b) is more similar to that of the observed in November than December, except for larger openings than observed along the southern part of the Victoria Land coast and in the Terra Nova Bay (Figure 30d). The observed low sea ice concentration in the northeastern part of the domain is not reproduced by the model, possible due to lack in melting and slow sea ice drift. Simulated and satellite-derived sea ice motion, where observed sea ice exists, compare well. Simulated sea ice thickness decreases in December over the whole grid except for the formation of a tongue of accumulated sea ice close to the ice shelf edge between 160°W and 170°W (Figure 30f).

In November of the second year of simulation (2001), simulated and observed coastal polynya openings and regions of relatively lower sea ice concentration match closely (Figure 31a, c). Simulated sea ice thickness to the north of the Ross Sea polynya area, near the Pennell Bank region, where observed lower sea ice concentration is not reproduced by the model, is low (Figure 31e). The similarity between the simulated
and observed sea ice cover continues into December, 2001 (Figure 31b, d) while simulated thinning of sea ice is domain-wide (Figure 31f). Unlike the first year of the simulation, both summer months in 2001 produce accumulated thick sea ice on and to the east of the western open boundary due to the reversal of the wind pattern supported by the observed sea ice drift.

Spatial extent of the openwater area over the polynya regions is relatively better simulated for 2001 than 2000 (Figures 30 and 31). Observed melting over the northern part of the domain in December due to ocean surface warming is not captured by the model, possibly due to ocean boundary conditions being climatological. Similar to winter months, simulated sea ice thickness provides a proxy for low sea ice cover in places where the simulated sea ice concentrations are overestimated.
III.2 REGIONAL ANALYSIS AND POLYNYA DYNAMICS

III.2.1 Regional Water Column Properties

The regional and seasonal variability of the temperature-salinity distribution in the Ross Sea is driven by atmospheric heating and cooling, continental shelf and ice shelf circulation, and the sea ice cycle. In March, at the end of austral summer, of the second year of simulation (2001), seawater in the ice shelf cavity is cold in the 34-34.8 salinity range with minimum temperatures well below the surface freezing point (Figure 32a). Simulated properties on the western Ross Sea shelf (see Figure 14a) are confined to a colder and more saline range compared to the eastern shelf (Figure 32b, c). Highest salinities reach 34.9 on the western shelf while the influx of ice shelf water onto the eastern shelf is evident by water at subfreezing temperatures. The Ross Sea continental slope region is comprised of warmer and fresher water masses, lacking cold and salty properties which occur on the continental shelf due to the ocean-sea ice-ice shelf interaction (Figure 32d). In March, the Antarctic Surface Water (AASW) is warmer and fresher over the continental shelf and slope due to the melting of sea ice and subsequent atmospheric heating. Warmer and fresher water entering the ice shelf
cavity from the western part of the Ross Sea enhances basal melting of the ice shelf. The saltiest end-member of the HISSW in the western shelf freshens while mixing with Lower Circumpolar Deep Water (LCDW) to form Modified Shelf Water (MSW).

In August, during the austral winter, AASW is cooled and becomes saltier on the continental shelf and slope, but not underneath the ice shelf due to lack of brine rejection from sea ice formation at the surface (Figure 33a). Over the western shelf, the region of salty, dense water formation, the winter salinity range is limited to 34.4-35 (Figure 33b). The ice shelf water flowing onto the eastern shelf also has a smaller range of salinities (Figure 33c). Simulated properties of water masses other than AASW on the continental slope are not significantly altered in winter (Figure 33d).
Figure 33. Regional simulated potential temperature and salinity distribution in August 2001. (a) Ross Ice Shelf water, (b) western shelf, (c) eastern shelf, and (d) slope.

One-dimensional model response

In order to investigate the response of the coupled sea ice-ocean model to regional variability of the Ross Sea water column and the lateral circulation, simulations were carried out with the vertical time-dependent model variation, described in Section II.3.2.

Five locations of varying depth and geographical emphasis were selected for simulations with the z-t model: two locations in the Terra Nova Bay and the Ross Sea polynya area are representative of the western shelf, and the other three are located in the eastern shelf, the central continental shelf break, and the western continental slope (Figure 34). Simulations start in June with the initial temperature-salinity distribution shown in Figure 34 and run for 10 years. Forcing is for the year 2000 and was extracted for each location from the fields used for the three-dimensional study.
Out of all five simulations, the z-t model reproduced the seasonal sea ice cycle only for the location on the western slope (Figure 35). The water column at initialization for this location is well-stratified and strong deep-nudging of simulated properties to initial values are done at temperatures higher than -1°C and salinities higher than 34.5. The location on the central shelf break is similarly well-stratified, however, the
thermocline is shallower compared to the location on the western slope. Consequently, nudging is done at temperatures higher than 0°C resulting in too much heat flux into the mixed layer from below to sustain the sea ice cycle. Simulated sea ice cover for shallower locations on the shelf were permanent and very thick. For the locations on the Ross Sea polynya area and the Terra Nova Bay, well-mixed water column at
Table 5. Comparison between the two-year time series of regional mean simulated and SSM/I sea ice concentration for the reference simulation

<table>
<thead>
<tr>
<th>Region</th>
<th>Mean deviation (%)</th>
<th>RMSD (%)</th>
<th>r</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ross Sea polynya (RSP)</td>
<td>-9.42</td>
<td>±19.78</td>
<td>0.89</td>
</tr>
<tr>
<td>Terra Nova Bay (TNB)</td>
<td>-9.38</td>
<td>±18.39</td>
<td>0.82</td>
</tr>
<tr>
<td>Victoria Land shelf (VLS)</td>
<td>-13.38</td>
<td>±17.79</td>
<td>0.95</td>
</tr>
<tr>
<td>Western Ross shelf</td>
<td>-10.95</td>
<td>±14.76</td>
<td>0.94</td>
</tr>
<tr>
<td>Eastern Ross shelf</td>
<td>-8.74</td>
<td>±13.53</td>
<td>0.93</td>
</tr>
<tr>
<td>Slope</td>
<td>-6.45</td>
<td>±12.42</td>
<td>0.94</td>
</tr>
<tr>
<td>Deep</td>
<td>-5.17</td>
<td>±12.31</td>
<td>0.94</td>
</tr>
</tbody>
</table>

* Sees Figure 14a for locations.

the freezing point supplies no heat available for melting sea ice in the summer due to nudging. For the location on the eastern shelf, there is some supply of heat from below the pycnocline, which slightly moderates the simulated sea ice thickness compared to the locations in the Ross Sea polynya area and the Terra Nova Bay, however such flux cannot decrease the sea ice concentration enough for solar radiation to penetrate and supply heat to the ocean surface layer for further melting.

III.2.2 Regional Sea Ice-Ocean Dynamics

Regional mean simulated sea ice concentration values are closer to the SSM/I data in the eastern Ross Sea shelf and the open ocean, possibly due to the highly dynamic nature of the western Ross Sea shelf, especially the polynya regions (Table 5). Time series of regional mean simulated sea ice concentrations are well correlated to the satellite data with regional mean deviations between -5.17% and -13.38%, and root-mean-squared deviations in the ±12.31%-±19.78% range. Victoria Land shelf and the Ross Sea and Terra Nova Bay polynya areas have relatively higher magnitudes of deviation compared to the rest of the domain.

Simulated mean thickness of sea ice formed due to thermodynamic processes in the Terra Nova Bay and the Ross Sea polynya area is 9 and 4 m thick per year, respectively, with very little local melting (Figure 36a). Regional averages for the western and eastern Ross Sea shelves and the Victoria Land coast are about 2 m annually, whereas the slope and the deep regions produce a mean of less than 1 m sea ice annually. In terms of total sea ice production, the western Ross Sea shelf produces as much sea ice as the deep Ross Sea (i.e. the region deeper than 3000 m), although the latter is about 2.5 times larger in area (Figure 36b and Table 6). The
eastern Ross Sea shelf also produces comparable amounts of sea ice to that of the western part. The Ross Sea polynya area produces a comparable amount of sea ice to the Victoria Land shelf, which is more than double the area. The Terra Nova Bay, occupying half a percent of the domain area, contributes about 4% of overall sea ice production. The Ross Sea polynya, on the other hand, occupies about 4.4% of the domain area while contributing about 14% to sea ice production (Table 6). The respective contributions of the Terra Nova Bay and Ross Sea polynya regions to sea ice production over the shelf only are 22.5 and 6% while occupying 1.5 and 12.5% of the area, respectively. Over the region delineated with the 1000 m isobath, there is only about 10-15% local melting of the thermodynamically formed sea ice, whereas close to 50% of thermodynamically formed sea ice melts locally over the slope and the deep regions (Table 6). 62.5% of all sea ice is formed on the continental shelf which is 35% of the domain area (Table 6).

Simulated sea ice area (volume) net export, calculated as the product of sea ice velocity and area (volume) at the regional boundaries, is a measure of freshwater flux out of the regions as well as an indication of sea ice divergence in the same area. Sea ice area net export is defined as the volume of sea ice exported at 1 m thickness per ice area. RSP is a region of net sea ice export except for a few net import episodes between December and February in the austral summer (Figure 37a). Furthermore, sea ice area export is consistently greater or equal to the volume export in winter, indicating that the thickness of the exported sea ice per ice area is 1 m thick or less. Total net sea ice export out of RSP is 450 km$^3$ over two years, equal to the thermodynamically formed sea ice in the region (Figure 37b, cf. Figure 36b). TNB, also a region of net export of thin, first-year ice (Figure 37c), experiences a transport

### Table 6. Regional cumulative sea ice production, area and net thermodynamic sea ice production contributions

<table>
<thead>
<tr>
<th>Region</th>
<th>2-yr cumulative production (m)</th>
<th>% area</th>
<th>% of domain total production</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ross Sea polynya (RSP)</td>
<td>8</td>
<td>4.5</td>
<td>10</td>
</tr>
<tr>
<td>Terra Nova Bay (TNB)</td>
<td>18</td>
<td>0.5</td>
<td>3</td>
</tr>
<tr>
<td>Victoria Land shelf (VLS)</td>
<td>5</td>
<td>9</td>
<td>13</td>
</tr>
<tr>
<td>Western Shelf</td>
<td>4</td>
<td>19</td>
<td>24</td>
</tr>
<tr>
<td>Eastern Shelf</td>
<td>4</td>
<td>16</td>
<td>20</td>
</tr>
<tr>
<td>Slope</td>
<td>1</td>
<td>17</td>
<td>5</td>
</tr>
<tr>
<td>Deep</td>
<td>2</td>
<td>48</td>
<td>25</td>
</tr>
</tbody>
</table>
Figure 36. Regional thermodynamic sea ice production. (a) Area mean thickness and (b) total volume. RSP, TNB, and VLS stand for Ross Sea polynya, Terra Nova Bay, and Victoria Land shelf, respectively. See Figure 14a for locations.

of the equivalent of thermodynamically formed total sea ice, about 120 km$^3$ sea ice over two years (Figure 37d). VLS sea ice exported is virtually equal to 1 m (Figure 37e) and total net export is slightly higher than the amount of sea ice formed in the region (Figure 37f). A few episodes of volume export in March, May, and August in 2000 show area import, meaning that thicker ice in the region is being replaced by thinner ice at higher area concentration. This is to be expected due to the export of accumulated thick, multi-year ice along the coast to the south of Cape Adare.

Vertically integrated lateral heat and salt fluxes into the regions is obtained by adding the horizontal advective and diffusive flux components throughout the water column at the region boundaries. When converted to a potential to melt meters of ice, vertically integrated lateral heat flux is a hypothetical indicator of the possible maximum contribution of sensible heat available to regulate or moderate sea ice thermodynamics. If all the lateral heat flux into the regions were used to melt sea ice, the contribution to melting over the two-year period would approximately be 80, 30, and 100% for RSP, TNB, and VLS, respectively (Figure 38a, cf. Figure 36a). All three regions experience net heat gain laterally, however, this calculation does not specify
at which depth the heat gain occurs. Salt flux, when integrated, is out of all three regions (Figure 38b), increasing in winter months during the freezing season. Mean simulated salt production in the Terra Nova Bay polynya area is the highest at about two times that of the Ross Sea polynya area and about four times the mean for the Victoria Land shelf.

III.2.3 Polynya Dynamics

Observed winter sea ice concentration for the RSP region is highly variable, ranging between 50 and 90%, whereas the simulated sea ice area fraction in winter is more
Figure 38. Regional mean lateral vertically integrated heat and salt fluxes. Lateral (a) time-integrated heat flux in terms of potential to melt sea ice and (b) salt production for Ross Sea polynya area (solid), Terra Nova Bay (dotted), and Victoria Land Shelf (dash-dot). Positive values denote flux into the region.

compact, in the 85-95% range (Figure 39a). Simulated sea ice thickness, which is less than 0.5 m a majority of the time, however, reflects almost all major polynya events observed in both winters. A comparison of the total area of less than 50% sea ice cover shows that the simulated timing of the polynya events is in agreement with observations; however, the polynya area is underestimated (Figure 39b). Polynya events coincide with increased wind speed (Figure 39c) and warmer peaks in air temperature (Figure 39d).

Simulated winter sea ice mass balance for the Ross Sea polynya area, as indicated by mean sea ice thickness (Figure 40a), is maintained by congelation and frazil ice formation (Figure 40b) and net export (Figure 40c). Sea ice loss due to basal melting and sublimation is minimal throughout the winter. Congelation and frazil ice formation rates are about 2 and < 0.5 cm d$^{-1}$, respectively, showing alternating peaks. Congelation ice formation increases with dropping air temperature (Figure 39d), which in turn causes a decrease in frazil ice formation due to increased insulation. Frazil ice formation, on the other hand, increases when open water area is greater (Figure 39a), sea ice is thinner (Figure 40a), sea ice export is greater (Figure 40c), and congelation ice formation is lower (Figure 40b). Snow-ice formation (Figure 40b) is near zero due to a very thin (less than 5 cm) layer of snow on sea ice (Figure 40a).

In austral summer, sea ice formation ceases (Figure 40b), sea ice concentration decreases rapidly (Figure 40a), and the sea ice mass balance is maintained by basal
Figure 39. Ross Sea polynya mean sea ice states, polynya area, and forcing for the reference simulation. (a) Simulated (black) and SSM/I (red) sea ice concentration, simulated sea ice thickness (dash dot), (b) simulated (black) and SSM/I (red) total area of less than 50% sea ice cover, and (c) wind and (d) air temperature forcing.  

Melting (Figure 40b) and net import into the region (Figure 40c), due to diminishing winds often reversed in direction (Figure 39c). During austral summer, the water column gains oceanic heat laterally from out of the region (Figure 40d, e) and solar heat at the ice-free ocean surface (Figure 40e). Lateral oceanic heat flux is into the region (Figure 40d), relatively higher during austral summer (Figure 40e). Net heat flux at the ocean surface (sum of open water-atmosphere heat flux, sea ice-ocean heat flux, and solar heat flux passing through sea ice to the ocean) is balanced by heat gain from frazil ice formation (i.e. restoration of
Figure 40. Ross Sea polynya area mean sea ice states, formation and net export rates, and ocean heat budget for the reference simulation. (a) Simulated sea ice concentration (solid) and thickness (dotted), and simulated snow thickness (dashed), (b) formation rate, (c) net export rate, (d) sea ice melt rate equivalent of vertically integrated lateral heat flux, and (e) regional ocean heat budget. Components of sea ice growth are congelation (black), frazil (red), and snow-ice (green) formation, evaporative flux from sublimation/condensation (blue), top (cyan), basal (magenta), and lateral (orange) melt. Heat budget terms are vertically integrated lateral heat flux (red), heat flux from frazil ice conversion (blue), and surface heat flux (green); black line denotes the sum.
the temperature of supercooled water to freezing) and lateral heat flux (Figure 40e),
the latter contributing about 90% percent of overall gain over two years of simulation.

The vertically integrated lateral heat flux calculation does not specify the depth
range at which the heat exchange between the polynya area and the surrounding
waters takes place. Meridional cross sections of simulated potential temperature

Figure 41. Simulated potential temperature cross sections along 180° (looking east)
from the reference simulation. Snapshots are for model days August 15 (a) 2000, (c)
2001 and December 23 (b) 2000, (d) 2001.
along the dateline show the depth distribution and seasonal variation of the oceanic sensible heat source for the region (Figure 41). In late austral winter, heat supplied by the CDW and entrained onto the continental shelf reaches the proximity of the polynya area south of Pennell Bank with a relatively warmer water mass occupying the 100-300 m depth range (Figure 41a, c). During the same period, the polynya area water column is well-mixed by deep convection driven by brine release due to sea ice formation (Figure 42a, c).
In early austral summer, the polynya area receives additional heat from solar radiation over the ice-free ocean surface (Figure 41b, d). Stratification increases due to cessation of brine release (Figure 42b, d) while cold water masses, partially sustained by flow from underneath the ice shelf, continue to occupy the deeper parts (Figure 41b, d).

Simulated winter sea ice concentration for the TNB region is in better agreement with observations compared to those for the RSP region (Figure 43a). Simulated sea ice thickness ranges between 0.15 and 0.40 m, considerably lower than the 0.3-0.6 m range for the RSP region, and reproduces most of the variability seen in the observed sea ice concentration field. Simulated area of less than 50% sea ice concentration compares more favorably to observations for the TNB region compared to the RSP. However, timing of these events do not match as closely (Figure 43b). Wind speed (Figure 43c) and warmer air temperature coincide with polynya events, but do not produce this effect in the first half of winter 2000 and the second half of winter 2001 (Figure 43d).

Simulated winter sea ice mass balance in the Terra Nova Bay is maintained by highly variable formation and export rates (Figure 44b, c). Total production from congelation and frazil ice formation reaches a maximum of about 10 cm d\(^{-1}\) in August. There is intermittent contribution to winter production from snow-ice formation depending on the height of the freeboard (Figure 44a). In January of both years of simulation, a period of relatively high rate of ice import into the region is followed in February by net ice export and basal melting (Figure 44c). Other than these two episodes, local melting in the region is near zero (Figure 44b). Lateral oceanic heat flux is comparable to that for the Ross Sea polynya area (Figure 44d, e), however the net heat loss at the ocean surface is greater. Net heat gain from frazil ice conversion in the area is about equal to sensible heat gain by lateral exchange (Figure 44e). As opposed to the Ross Sea polynya area, where the surface heat loss is balanced by the sensible heat gain from lateral fluxes and frazil ice formation (Figure 40e), the Terra Nova Bay polynya water column experiences about \(2 \times 10^8 \text{ J m}^{-2}\) surplus warming over two years of simulation.

### III.3 MODEL SENSITIVITY TO TERRA NOVA BAY WIND FORCING

In order to investigate the sensitivity of the simulated polynya processes in the Terra Nova Bay area, the coupled model is forced by the ECMWF ERA-40 winds
Figure 43. As in Figure 39, but for the Terra Nova Bay polynya.

(Figure 13a) unmodified for the region (“ECMWF” simulation). The results thus obtained are evaluated with respect to sea ice dynamics in the polynya area and its effects on the ocean circulation and water mass distribution.

### III.3.1 Effects on Sea Ice Dynamics

Compared to the results from the reference simulation (Section III.1), the ECMWF simulation yields sea ice concentrations slightly closer to the observed values over the Victoria Land shelf and the western Ross Sea in terms of mean and magnitude (Table 7). For other regions, the reference simulation is in better agreement with the satellite data.
Simulated mean thermodynamic sea ice production in the Terra Nova Bay with ECMWF wind forcing is 8 m over two years, which is less than half compared to the 17.5 m produced by the simulations forced with the AWS data (Figure 45). Similar to the reference case, the simulated sea ice that is exported from the Terra Nova Bay is thin and except for the summer 2000-01, during which sea ice is imported into the region, is consistently out of the bay area (Figure 46a). Sea ice import continues...
Table 7. Comparison between the two-year time series of regional mean simulated and observed sea ice concentration for the ECMWF simulation

<table>
<thead>
<tr>
<th>Region</th>
<th>Mean deviation (%)</th>
<th>RMSD (%)</th>
<th>r</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ross Sea polynya (RSP)</td>
<td>-10.19</td>
<td>±20.61</td>
<td>0.89</td>
</tr>
<tr>
<td>Terra Nova Bay (TNB)</td>
<td>-11.52</td>
<td>±20.11</td>
<td>0.60</td>
</tr>
<tr>
<td>Victoria Land shelf (VLS)</td>
<td>-10.62</td>
<td>±15.77</td>
<td>0.94</td>
</tr>
<tr>
<td>Western Ross shelf</td>
<td>-10.33</td>
<td>±14.30</td>
<td>0.94</td>
</tr>
</tbody>
</table>

* See Figure 14a for locations.

through both summers. A total of 60 km$^3$, about 5 km$^3$ in excess of local production, of sea ice is exported out of the TNB region by the ECMWF simulation over two years as opposed to 120 km$^3$ by the reference case (Figure 46b, cf. Figure 37d).

Figure 45. Wind forcing effect on the mean thermodynamic sea ice production in the Terra Nova Bay. Reference (solid) and ECMWF (dotted) simulations.

Terra Nova Bay simulated winter sea ice concentration for the ECMWF simulation compares better to the satellite data, especially in the second year of the simulation, than that for the reference simulation (Figure 47a). Winter sea ice thickness for the ECMWF simulation is in the first winter thicker than and in the second winter comparable to that of the reference simulation. Elevated thickness over summer 2000-01 exceeds any winter value for both years. Snow layer on sea ice is also thicker in the first 15 months. The ECMWF simulation overestimates summer sea ice cover in both years, possibly due to a seasonal reversal in the direction of wind forcing (Figure 47c). There are more polynya events in the ECMWF simulation, yielding larger polynya areas, lasting longer, and in better agreement of timing with the observed data (Figure 47b). Observed polynya events in December 2000, May and November 2001 are especially well captured by this simulation, and simulated polynya events
Figure 46. Regional net sea ice area and volume export for the Terra Nova Bay for the ECMWF simulation. Flux rate (a) and time-integrated (b) area (dotted) and volume (solid) net export are shown. For consistency in units, sea ice area export is expressed as the equivalent volume export of 1 m thick sea ice.

better reflect the warmer peaks in air temperature forcing (Figure 47d).

For the ECMWF simulation, congelation and frazil ice formation rates are significantly lower compared to the reference case (Figure 48b). Although small, there is more snow-ice formation in both winters due to greater snow accumulation on sea ice. Import of sea ice into the region and consequent basal melting in December continues through February (Figure 48b, c). In response to low ECMWF winds, overall export of ice is slower compared to the reference simulation (Figure 48c), whereas oceanic sensible heat into the region is greater, increasing during the second year of the simulation (Figure 48d). Lateral oceanic heat flux provides about a factor of 8 times more than that from frazil ice formation (Figure 48e). Such a ratio between two heat sources renders the Terra Nova Bay polynya to behave like the Ross Sea polynya when forced with ECMWF winds. Also for this simulation, the Terra Nova Bay polynya region experiences a greater warming tendency (i.e. about $10 \times 10^8$ J m$^{-2}$) compared to the reference case.

III.3.2 Effects on Water Column Properties

Two-year total vertically integrated lateral heat flux into the TNB region for the ECMWF simulation is more than two times greater than that for the reference simulation (Figure 49a). This flux is equivalent to heat that can melt 14 m of sea ice, more than the 8 m mean thermodynamically produced sea ice in the region. Lateral salt flux for both cases, is out of the region when integrated, however, is significantly less for the ECMWF simulation throughout the winter of 2000 and late winter 2001 (Figure 49b).

West-east cross sections along the western 200 km of the TNB-RSP section (see
Figure 47. As in Figure 43, but for the ECMWF simulation.

Figure 14b) of simulated potential temperature show the effects of sea ice formation rates on the thermal structure of the Terra Nova Bay water column in late winter (Figure 50). For the reference simulation, warmer water masses do not intrude into the well-mixed Terra Nova Bay polynya area, increasing the potential to form frazil ice throughout the water column (Figure 50a, c). For the ECMWF simulation, on the other hand, the brine release from moderated sea ice formation is not sufficient for deep convection, and warmer water intrudes the polynya area, further decreasing the potential to form sea ice (Figure 50b, d).

North-south salinity cross sections along the Victoria Land shelf in the western Ross Sea (WRS section; see Figure 14b) show the difference in the northern extent
Figure 48. As in Figure 44, but for the ECMWF simulation.

of the HSSW resulting from different Terra Nova Bay wind forcing (Figure 51). In February of the second year of simulations, the shelf water that occupies the deeper troughs below 200 m south of the continental slope are up to 0.25 more saline for the reference simulation.

The eastern extent of the effects of Terra Nova Bay winds on the production of dense and salty shelf water, on the other hand, is not as far reaching as the northern as
Figure 49. Wind forcing effect on regional mean vertically integrated lateral heat and salt fluxes in the Terra Nova Bay. Lateral (a) time-integrated heat influx in terms of potential to melt sea ice and (b) salt outflux for the reference (solid) and ECMWF (dotted) simulations.

indicated by the simulated west-east salinity cross sections (Figure 52) that span the Terra Nova Bay and the Ross Sea polynya regions (i.e. farther east along the TNB-RSP section). In February of the first year of simulation, the reference simulation produces HSSW with salinity in the 34.81-34.86 range, filling the western trough below 200 m, whereas this water mass is not created by the ECMWF simulation (Figure 52a, b). In February of the second year of simulation, the ECMWF simulation produces water in this salinity range, however limited to only the bottom 200 m of the trough (Figure 52d). By the same time, the bottom 400 m in the same location is occupied by HSSW of salinity 34.86-34.90 for the reference simulation (Figure 52c). For both simulations, the model reproduces the observed west-east salinity gradient over the Ross Sea continental shelf [Jacobs and Giulivi, 1998, 1999], however, the ECMWF simulation fails the produce the saltier end-member of HSSW.

Horizontal cross sections of simulated yearly mean salinity at 400 and 500 m show the effect of Terra Nova Bay winds on the formation of salty, dense water in western Ross Sea and the redistribution of this water mass (Figures 53-56). At 400 m, the yearly mean modelled salinity field from the reference simulation along the Victoria Land shelf is between 34.68 and 34.88 with the northern extent north of 74°S in 2000 (Figure 53a) and reaching further north close to the shelf break the following year
Figure 50. Simulated potential temperature cross sections along the western 200 km of the TNB-RSP section. Snapshots are for August 15, 2000 (top row) and 2001 (bottom row) from the reference (a, c) and ECMWF (b, d) simulations. See Figure 14b for the location of the section.

(Figure 53b). Banded salinity structure off the shelf break results from the advection of boundary conditions by the enhanced slope current (Figure 24). The ECMWF simulation fails to produce the salty water over the western shelf, yielding maximum salinities less than 34.85 with the northern extent south of 74°S in 2000 (Figure 54a), retreating southward in 2001 (Figure 54b). However, the salinity field to the north
Figure 51. Simulated salinity and potential temperature cross sections along the WRS section. Snapshots are for February 1 2001 from the (a, b) reference and (c, d) ECMWF simulations. See Figure 14b for the location of the section.

and northeast of Ross Island and to the east of 175°W on the shelf near Cape Colbeck is slightly more saline for the ECMWF simulation compared to the reference case.

For the reference case, at 500 m, the yearly mean modelled salinity field to the north of Drygalski Ice Tongue is uniformly salty, in the 34.85-34.88 range (Figure 55a), spreading east of 160°E and filling the northern trough south of Cape Adare in
Figure 52. Simulated salinity cross sections along the TNB-RSP section. Snapshots are for February 11, 2000 (top row) and February 1, 2001 (bottom row) from the reference (a, c) and ECMWF (b, d) simulations. See Figure 14b for the location of the section.

Over the same region, salinities are in the 34.78-34.83 range for the ECMWF simulation with the northern extent of the water mass at 74°S in 2000 (Figure 56a). The salt content along the Victoria Land shelf is less in the second year of simulation, while increasing to the west of Cape Colbeck (Figure 56b).

Horizontal cross sections of simulated yearly mean potential temperature at 300 m
Figure 53. Yearly mean model salinity at 400 m for the reference simulation. (a) 2000 and (b) 2001. Bathymetry contour shown is the 1000 m isobath.

Figure 54. As in Figure 53, but for the ECMWF simulation.
Figure 55. Yearly mean model salinity at 500 m for the reference simulation. (a) 2000 and (b) 2001. Bathymetry contour shown is the 1000 m isobath.

Figure 56. As in Figure 55, but for the ECMWF simulation.
Figure 57. Yearly mean model potential temperature at 300 m for the reference simulation. (a) 2000 and (b) 2001. Potential temperature values greater than 0°C are represented by the same contour level. Bathymetry contour shown is the 1000 m isobath.

Figure 58. As in Figure 57, but for the ECMWF simulation.
show the locations along the shelf break of warmer Circumpolar Deep Water (CDW) intrusions onto the continental shelf at intermediate depths as well as the extent of the influence of such intrusions (Figures 57-58). Warmer water intrusions take place to the north and west of the Pennell Bank extending between Cape Adare and 180°, and over most of the eastern shelf except to the west of Cape Colbeck (Figure 57a). The influence of intrusions over the northwestern Ross Sea reaches as far south as 76°S along 160°E. While the warmer water mass spreads to the east of this location into 2001, a colder tongue of water along the coast moves slightly northward towards Cape Adare (Figure 57b). Main difference between the reference simulation and the ECMWF case is to the north of 74°S along the Victoria Land coast where water at intermediate depth for the latter is warmer in 2000 (Figure 58a) and continues to warm to the south of that latitude, reaching the coast of the Terra Nova Bay in 2001 (Figure 58b). Despite the warming along the coast, the ECMWF simulation retains colder water mass to the east of 160°E in the northern part of the Ross Sea polynya area.

III.3.3 Effects on Water Mass Transformations

The effects of Terra Nova Bay on the thermohaline structure resulting from local sea ice dynamics influence through circulation the water mass transformations over the western Ross Sea shelf and within the ice shelf cavity. Simulated mean ice shelf basal melt rate is consistently about 1-2 cm yr⁻¹ greater for the ECMWF simulation throughout the most part of both winters of simulation and in February 2000, while slightly less in February 2001 (Figure 59).

Figure 59. Simulated mean ice shelf melt rate for the reference (solid) and ECMWF (dotted) simulations.
In order to quantify the simulated circulation of various water masses, vertically-integrated annual mean potential temperature-salinity (Θ-S) distribution of total flow in and out of the selected regions and sections is computed. Bulk of the flow southward across the RIS section (see Figure 14b) is aligned on the surface freezing point line for both simulations (Figures 60 and 61). For the reference simulation, flux out of the ice shelf region in the first year is concentrated at magnitudes between 60 and 150 mSv, in the 34.60-34.75 salinity range with potential temperatures between -2.1°C and the surface freezing point (Figure 60a). The flux into the ice shelf region, similar in magnitude and at around the surface freezing point, has salinities in the 34.75-34.80 range. Total exchange of shelf water types is relatively smaller in magnitude in the second year of the reference simulation, while slow influx of surface and circumpolar deep water types are scattered over a wider range of potential temperature and salinity (Figure 60b). The results from the ECMWF simulation have a similar pattern of water mass exchange, although part of the high salinity shelf water flux into the region is slightly saltier in the first year of the simulation (Figure 61a). The ECMWF case also yields a decrease in the magnitude of overall exchange of shelf water types from the first simulation year to the next, however, over two years of simulation, total HSSW flow into the region is greater for the ECMWF simulation compared to that for the reference simulation (Figure 61b). The ECMWF simulation also yields an increase in the Θ-S range of the surface and circumpolar deep water types flowing into the iceshelf cavity in the second year of the simulation, which may be the reason for greater mean ice shelf melt during that period (Figure 59).

The Θ-S distribution of flux into and out of the PLY region (see Figure 14b) shows that the cold, saltier water, in the 34.74-34.77 salinity range at the surface freezing point, flowing out of the region is replaced by the warmer and less salty AASW in both years of the reference simulation (Figure 62). Magnitude of the outflux of dense, salty water is between 80 and 200 mSv in the first year (Figure 62a), decreasing in the second (Figure 62b). In the first year of the ECMWF simulation, the outflux is larger in magnitude compared to the reference simulation (Figure 63a), however, it is significantly diminished in the second simulation year (Figure 63b). It should be noted that the south edge of the PLY region includes exchange at western part of Ross Ice Shelf edge as shown by inflow of water with temperature well below the surface freezing point.

The Θ-S distribution of flux across the SHB section (see Figure 14b) show the characteristics of the water masses exchanged between the shelf and the deep ocean over the Ross Sea continental margin (Figures 64 and 65). For the reference simulation,
Figure 60. Simulated yearly mean vertically-integrated Θ-S volume distribution of net flow into/out of the Ross Ice Shelf cavity (RIS section) from the reference simulation for (a) 2000 and (b) 2001. Negative values denote flux southward into the RIS section.

Figure 61. As in Figure 60, but for the ECMWF simulation.
Figure 62. As in Figure 60, but for Terra Nova Bay and Ross Sea polynya regions (PLY region) from the reference simulation.

Figure 63. As in Figure 62, but for the ECMWF simulation.
while the inflow onto the continental shelf is concentrated at the LCDW range, water masses that flow off the shelf break onto the continental slope include MSW, new AABW, the colder components of AASW, LSSW, and MCDW, and the less saline component of HSSW. Outflux of AABW somewhat decreases in the second year of the simulation, whereas the Θ-S range of MSW outflux increases (Figure 64b). The ECMWF simulation shows a similar pattern of exchange with the main difference being decreased outflow and Θ-S range of MSW (Figure 65).

Yearly mean net volume flux of deep water masses show that MSW is abundant over the western half of the Ross Sea continental shelf, flowing northward downslope to the east of 180° between 80°S and 75°S with magnitudes up to about 3 Sv and westward along the shelf break with magnitudes less than 1 Sv (Figure 66a, d). Along the Victoria Land shelf, MSW flows northward along the coast (better developed in 2001) and southward offshore, the latter being larger in magnitude, in the vicinity of 1 Sv. The southward transport continues along the western part of the ice shelf edge and into the cavity at about the same magnitude.

HSSW transport also takes place predominantly in the western Ross Sea with smaller signals along the ice shelf edge with an eastern extent at around 160°W (Figure 66b, e). East of 180° transport of HSSW has smaller magnitudes than that of the MSW and is limited to higher latitudes. The fluxes of HSSW downslope towards the shelf break between 180° and 160°E, into the ice shelf cavity to the south of Ross Sea polynya area, and downslope towards Cape Adare along the Victoria Land coast are comparable in magnitude to those of the HSSW. Net flux of LSSW, on the other hand, occurs over the eastern shelf, to the east of 180°, downslope and less than 1 Sv in magnitude (Figure 66c, f). Major differences from the first year of the simulation to the next are the strengthening transport of MSW northward along the Victoria Land coast, weakening transport of HSSW into the ice shelf cavity in the second year, and weakening transport of LSSW northward to the east of 180°.

The spatial and magnitude distribution of MSW transport for the ECMWF case is different from that of the reference simulation offshore of Victoria Land and over the southern part of the Ross Sea polynya area (Figure 67a, d). MSW is transported southward along both flanks of the 500 m isobath along the Victoria Land coast in the first year and the offshore component diminishes in the second year. The eastward transport over the southern part of the Ross Sea polynya area produced by the reference simulation also is not developed in the ECMWF case. The influence of sea ice formation rates in the Terra Nova Bay polynya area on maintaining the circulation of HSSW off the coast of Victoria Land is indicated by the absence of
Figure 64. As in Figure 60, but for the Ross Sea shelf break (SHB section).

Figure 65. As in Figure 64, but for the ECMWF simulation.
northward transport of this water mass along the western boundary (cf. Figure 66b, e and 67b, e). HSSW net flux to the north of 76°S is virtually absent, revealing a more significant difference between the two simulations. Furthermore, the flux into the ice shelf cavity is greater in the first year and the direction of transport is reversed in the second year of the ECMWF simulation. LSSW flux, on the hand, does not yield a significant difference between the two simulations (Figure 67c, f).

Time series of net volume outflux of various deep water masses out of selected regions and along sections are computed to compare relative magnitudes and investigate seasonal variations. For the reference simulation, net LSSW outflux across the RIS section is balanced by net HSSW influx in austral winter and spring (April through December), and by net MSW influx in summer and fall (Figure 68a). Exchange of MCDW/LCDW across the section is minimal. Net inflow of HSSW and outflow of LSSW decrease in the second year of the simulation, whereas outflow of MSW increases (Table 8). Over two years, the ice shelf cavity acts effectively as a sink at the rate of 0.39 Sv for HSSW and 0.23 Sv for MSW, while is a source at 0.61 Sv for LSSW.

RSP region is characterized by outflux of HSSW May through January, reaching peaks of about 2 Sv in August 2000 and July 2001 (Figure 68b) with a mean of 0.64 Sv over two years (Table 8). During the same periods, MSW flows into the region with peaks of about 1 Sv and out of the region in early winter, yielding a net mean rate of 0.43 Sv. Net LSSW outflux is minimal at the rate of about 0.60 Sv in and out of the region. MCDW flux into the region is small, netting 0.08 over two years. The region acts as a sink for MCDW/LCDW and MSW, while producing HSSW.

HSSW outflux from the TNB region follows a similar temporal pattern to that of the RSP region with a two year mean net outflux at 0.14 Sv (Figure 68c, Table 8). MCDW, MSW, and LSSW net inflow into the region at small rates balances the HSSW outflux.

VLS region experiences a greater influx of MCDW/LCDW, while producing more HSSW than the RSP region (Figure 68d). During the winter months, net inflow of MCDW/LCDW, MSW/AABW, and LSSW are small, implying that AASW is the main source for HSSW production. Overall, the region is a source for HSSW and MSW, and a sink for MCDW/LCDW and LSSW (Table 8).

The PLY region, the extended area that covers most of the western Ross Sea shelf, is characterized more significant influx rates of MCDW/LCDW and LSSW and outflux of HSSW and MSW/AABW (Figure 68e). MCDW/LCDW consistently flows into the region throughout the year at a mean net influx of 0.73 Sv (Table 8). Net mean
Figure 66. Yearly mean net volume flux of water masses for the reference simulation. Fluxes are for the simulation years 2000 (left column) and 2001 (right column), water masses MSW/AABW (a, d), HSSW (b, e), and LSSW (c, f). Each arrow is representative of a 100×100 km area with its base at the center of the region. See Figure 15 for water mass definitions.
Figure 67. As in Figure 66, but for the ECMWF simulation.
Figure 68. Net volume outflux of water masses from the regions and sections analyzed for the reference simulation. Flux in and out of (a) RIS section, (b) RSP region, (c) TNB region, (d) VLS region, (e) PLY region, and (f) SHB section for MCDW/LCDW (red), MSW/AABW (green), HSSW (blue), and LSSW (cyan) are shown. See Figure 14 for locations, and Figure 15 for water mass definitions.
Table 8. Regional mean volume fluxes (in Sv) of deep water masses (influx/outflux/net) from the reference simulation

<table>
<thead>
<tr>
<th>Region</th>
<th>MCDW/LCDW</th>
<th>MSW/AABW</th>
<th>HSSW</th>
<th>LSSW</th>
</tr>
</thead>
<tbody>
<tr>
<td>RIS</td>
<td>0.02/0.01/-0.01</td>
<td>0.35/0.18/-0.17</td>
<td>3.32/2.74/-0.57</td>
<td>1.03/1.76/0.73</td>
</tr>
<tr>
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<td>0.04/0.03/-0.01</td>
<td>0.56/0.28/-0.28</td>
<td>2.46/2.25/-0.20</td>
<td>0.89/1.38/0.49</td>
</tr>
<tr>
<td></td>
<td>0.03/0.02/-0.01</td>
<td>0.46/0.23/-0.23</td>
<td>2.89/2.50/-0.39</td>
<td>0.96/1.57/0.61</td>
</tr>
<tr>
<td>RSP</td>
<td>0.20/0.15/-0.06</td>
<td>1.96/1.46/-0.50</td>
<td>2.26/2.92/0.66</td>
<td>0.69/0.66/-0.03</td>
</tr>
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<td>1.95/2.58/0.62</td>
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</tr>
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<td>0.61/0.60/-0.01</td>
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<td>0.88/1.02/0.14</td>
<td>0.10/0.07/-0.03</td>
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<td>0.09/0.03/-0.06</td>
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<td>0.39/0.35/-0.03</td>
<td>0.88/1.02/0.14</td>
<td>0.09/0.05/-0.04</td>
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<td>3.81/5.22/1.41</td>
<td>1.84/1.77/-0.08</td>
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<td>2.52/1.76/-0.76</td>
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<td>0.07/0.25/0.18</td>
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<td>2.77/4.98/2.21</td>
<td>0.02/0.11/0.09</td>
<td>0.05/0.18/0.13</td>
</tr>
<tr>
<td></td>
<td>9.97/9.28/-0.69</td>
<td>2.89/5.12/2.23</td>
<td>0.02/0.12/0.11</td>
<td>0.06/0.22/0.15</td>
</tr>
</tbody>
</table>

a See Figure 14 for locations.
b See Figure 15 for water mass definitions.
c For "RIS", outflux is towards the model north.
d For "SHB", outflux is towards the model north/east, i.e. off the shelf break.
e Rows denote the means for the first year, the second year, and two years combined, respectively.

LSSW transport into the region is 0.43 Sv, decreasing in summer. MSW/AABW is mainly produced in summer and fall, amounting to a net rate of 0.77 Sv over two years. The extended polynya region produces a mean net HSSW outflux of 1 Sv.

Along the section that approximately tracks the continental shelf break (SHB section; see Figure 14b), MSW/AABW is consistently exported with a net mean outflux of about 2.2 Sv (Figure 68f, Table 8). Net mean import of MCDW/LCDW takes place at 0.7 Sv, implying that AASW is an important source in the balance, since net HSSW and LSSW outflux across the section is small compared to the MSW/AABW-MCDW/LCDW exchange.
Figure 69. As in Figure 68, but for the ECMWF simulation.
Table 9. Regional mean volume fluxes (in Sv) of deep water masses (influx/outflux/net) from the ECMWF simulation

<table>
<thead>
<tr>
<th>Region&lt;sup&gt;a&lt;/sup&gt;</th>
<th>MCDW/LCDW&lt;sup&gt;b&lt;/sup&gt;</th>
<th>MSW/AABW</th>
<th>HSSW</th>
<th>LSSW</th>
</tr>
</thead>
<tbody>
<tr>
<td>RIS&lt;sup&gt;c&lt;/sup&gt;</td>
<td>0.01/ 0.00/ -0.01&lt;sup&gt;ε&lt;/sup&gt;</td>
<td>0.31/ 0.13/ -0.18</td>
<td>3.68/ 3.05/ -0.63</td>
<td>0.99/ 1.78/ 0.79</td>
</tr>
<tr>
<td></td>
<td>0.03/ 0.01/ -0.02</td>
<td>0.60/ 0.35/ -0.25</td>
<td>2.74/ 2.54/ -0.20</td>
<td>0.85/ 1.30/ 0.45</td>
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<tr>
<td></td>
<td>0.02/ 0.01/ -0.01</td>
<td>0.46/ 0.24/ -0.22</td>
<td>3.21/ 2.80/ -0.41</td>
<td>0.92/ 1.54/ 0.62</td>
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<td>RSP</td>
<td>0.19/ 0.14/ -0.05</td>
<td>1.69/ 1.15/ -0.54</td>
<td>2.71/ 3.50/ 0.79</td>
<td>0.80/ 0.67/ -0.13</td>
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<td>0.34/ 0.21/ -0.14</td>
<td>1.90/ 1.74/ -0.16</td>
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<td>0.60/ 0.60/ 0.00</td>
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<td>0.26/ 0.17/ -0.09</td>
<td>1.80/ 1.45/ -0.35</td>
<td>2.38/ 3.04/ 0.65</td>
<td>0.70/ 0.63/ -0.07</td>
</tr>
<tr>
<td>TNB</td>
<td>0.05/ 0.03/ -0.02</td>
<td>0.77/ 0.82/ 0.05</td>
<td>0.69/ 0.76/ 0.07</td>
<td>0.11/ 0.07/ -0.04</td>
</tr>
<tr>
<td></td>
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<td>0.24/ 0.25/ 0.02</td>
<td>0.14/ 0.13/ -0.01</td>
</tr>
<tr>
<td></td>
<td>0.09/ 0.05/ -0.04</td>
<td>0.97/ 1.05/ 0.08</td>
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<td>0.13/ 0.10/ -0.03</td>
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<td>1.64/ 1.32/ -0.32</td>
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<td>0.02/ 0.11/ 0.09</td>
<td>0.06/ 0.20/ 0.14</td>
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</table>

<sup>a</sup> See Figure 14 for locations.

<sup>b</sup> See Figure 15 for water mass definitions.

<sup>c</sup> For “RIS”, outflux is towards the model north.

<sup>d</sup> For “SHB”, outflux is towards the model north/east, i.e. off the shelf break.

<sup>ε</sup> Rows denote the means for the first year, the second year, and two years combined, respectively.

The effects of Terra Nova Bay winds on the simulated water mass transport rates can be identified in the time series of net volume flux out of the TNB region of deep water masses (Figure 69). From September through November of the first year of simulation, HSSW flux is into the region and diminishes to near zero afterwards. During the winter of the second year of simulation, outflux from the region is dominated by that of the AABW, at about the same magnitude as the previous year, while for the reference case overall dense water outflux decreases (Figure 68c). Mean net outflux of HSSW is insignificant at 0.01 Sv, however, AABW outflux, at 0.16 Sv, is almost twofold compared to that from the reference case. MSW flux, mean net outflux of
which was near zero for the reference case, is into the region at a mean 0.05 Sv. No notable differences in water mass flux exist between the two simulations along the south edge of the RIS region and for the RSP region (Table 9). Meanwhile, MSW and HSSW outfluxes are greater and LSSW influx is smaller by about 0.1 Sv for the PLY region. Along the SHB section, MCDW influx is increased while MSW and AABW outfluxes are decreased also by about 0.1 Sv. For the Terra Nova Bay, in particular, over two years of simulation, MCDW/LCDW and MSW/AABW gradually replaces HSSW in terms of net formation and volume transport under weaker wind forcing.
CHAPTER IV
DISCUSSION

The regional, mesoscale-resolution sea ice-ocean model with dynamic and thermodynamic sea ice and thermodynamic ice shelf components, as described in Section II.3.3, provides a framework to study the coupled sea ice-ice shelf-ocean system in a one-way interaction with the atmosphere in the Ross Sea. The results presented in Chapter III described the simulated sea ice and ocean states, fluxes, and motion with respect to their regional and seasonal distribution. The sections that follow discuss the model results with emphases on the Ross Sea seasonal sea ice cycle, polynya processes, and water mass distribution and transformations in an attempt to validate the simulated fields and place the current modeling effort in context with relevant previous studies.

IV.1 ROSS SEA SEASONAL SEA ICE CYCLE

IV.1.1 Sea Ice Concentration and Drift

Domain averaged simulated sea ice concentration, which also is a measure of the total sea ice area, is well correlated to the satellite data (Figure 17). The mean difference over two years between the simulated and observed sea ice concentration is within the general accuracy margin of the validation field (see Section II.3.3). The magnitude of the difference decreases and the correlation coefficient increases with a lag of 11 days. Improvement with a lagged correlation may be due the initialization of the sea ice cover on September 15 with the monthly mean field. Furthermore, Timmermann et al. [2002] argue that starting coupled simulations in summer with no initial sea ice cover may result in a short adjustment time. Initialization field, therefore, may explain part of the timing mismatch between the observed and simulated sea ice concentration fields for which the slopes of winter freezing and summer melting compare favorably.

The difference between two fields in terms of the length of ice season and compactness of the ice cover, on the other hand, requires an evaluation of the seasonal progress of horizontal distribution of simulated sea ice. Throughout the austral winter, simulated sea ice cover is consistently compact in the 85-100% range with minimal lead opening. Low concentrations resulting from divergent motion are limited to narrow regions along the Ross Ice Shelf front and the north of the Drygalski Ice Tongue.
(Figures 26-29(a, b)). Thorndike et al. [1975] describe the competition between sea ice thermodynamics and dynamics as the first seeking the mean and the latter the extremes. Although the dynamic model opens leads scattered to the north of the continental shelf break (i.e. the 1000 m isobath) at an average rate of up to about 4% d\(^{-1}\) between May and September in both years of the simulation (not shown), tendency of the thermodynamic model for lateral growth removes open water fraction by freezing.

The minimum thickness allowed in the thinnest sea ice category by the thermodynamic model is set to 0.01 m as the default value for CICE. For this study, a value of 0.1 m, as modified for the CSIM5 distribution, is used. This parameterization determines how fast the open water fraction decreases in leads and polynyas by frazil ice collecting laterally on the sides of consolidated pack ice, which is a process not well understood [Pease, 1987; Van Woert, 1999b]. Prescribed thickness used in sea ice and polynya modeling [e.g., Pease, 1987; Galleé, 1997] can be as high as 0.5 m. Methods for variable collection thickness as a function of frazil ice drift speed are also used Biggs et al. [2000]. Among other parameterizations that determine the lateral versus vertical distribution of newly formed sea ice, the minimum thickness (i.e. frazil ice collection thickness) may need to be better tuned to moderate compactness in winter. Nevertheless, simulated rate of lead opening due to dynamics, even if not counteracted on by thermodynamics, may still not be high enough to reproduce the observed sea ice concentration variability due to inaccuracies in the forcing field. The actual winter sea ice cover is inherently compact, and hence, the differences between the simulated and observed winter sea ice concentration are in the -5 to 15% range over most of the domain, except in coastal polynya regions (Figures 26-29(c, d)) which contain a high fraction of thin ice (Figures 26-29(e, f)).

In general, overestimating winter sea ice compactness within the accuracy margins of sea ice concentration derived from satellite microwave data is not uncommon in modeling studies using CICE/CSIM for regional [e.g., Hunke and Ackley, 2001; Prasad et al., 2005] and circumpolar studies [e.g., Holland et al., 2006] of Antarctica. Studies using different sea ice models in the coupled implementation for circumpolar domains have reported simulated Ross Sea winter sea ice concentrations variably as elevated [Fichefet and Morales Maqueda, 1997; Timmermann et al., 2002; Assmann, 2003; Wu et al., 2003] and favorably reproduced [Wu et al., 1997; Weatherly et al., 1998]. Sea ice concentration from two Ross Sea regional modeling studies, reported only for November, is higher than climatological SSM/I data for that month [Fichefet and Goosse, 1999; Reddy et al., 2007]. Although model thermodynamics and dynamics
are formulated differently, common to most of the mentioned sea ice modeling work is the atmospheric forcing fields adapted or derived from low-resolution reanalysis products.

Simulated summer values explain yearly mean discrepancies of greater magnitude, which result from the zonal bias of wind forcing south of the northern open boundary and the inaccuracy of the imposed sea ice states at the eastern open boundary. Deflection of winds near the northern open boundary on a yearly mean basis (Figure 21b, e) is not apparent from the observed sea ice drift (Figure 18b, d), which can serve as a proxy for the actual wind field. Sea ice accumulates on the northern part of the domain throughout the winter and the model cannot remove such excess in time, resulting in longer ice seasons. Over the eastern part of the domain, on the other hand, the yearly mean sea ice concentration is underestimated not because of excessive melting (Figure 22a, c), but inaccurate representation of sea ice drifting westward from the Amundsen Sea. Assmann et al. [2005] reported positive correlation in model results between the northward export of sea ice from western Ross Sea and the intrusion of Amundsen Sea ice. Imposed monthly climatological sea ice fields on the eastern boundary disrupts this coupling. Climatological sea ice concentration at a fixed thickness drifting westward into the domain does not provide enough volume, which in turn, results in the modelled underestimates of sea ice concentration over the eastern part of the domain.

Other regions where winter accumulation is not balanced by summer removal (Figure 19) are to the north and east of the Terra Nova Bay and on the eastern shelf between 165°W and 170°W. Both regions show simulated thicknesses greater than 0.5 m (Figure 20) and experience convergence due to slowing zonal sea ice drift downstream (Figure 21a, d). Therefore, most of the contribution to the annual mean sea ice concentration overestimate for these regions comes from the summer values. For the Ross Sea polynya area; however, the difference between yearly mean simulated and observed sea ice concentration decreases due to a better reproduced summer polynya.

Simulated yearly mean sea ice drift for this study is mainly wind-driven, but is also affected by relatively fast ocean surface flow (Figure 21). Consolidated winter ice pack is not sensitive to surface currents and qualitatively compares better to the observed drift. The drift responds year-round to ocean surface flow along 180° on the shelf, over the continental slope, and offshore along the abyssal current flowing northeastward. During the summer, however, the contribution of ocean surface flow to the dynamics of breaking pack ice over the continental shelf is greater (Figures 39c,
30d, 31d) due to the weakening wind field (e.g., Figures 39c and 43c; also cf. Figures 2d, 3d in Budillon et al. [2000]). This results in excess ice accumulation on the east and west sides of the expanding edge of the summer Ross Sea polynya. During the preliminary part of the current study, simulations forced by the NCEP/NCAR wind field did not show ice accumulation over the western shelf between 74°S and 76°S (not shown). NCEP/NCAR wind forcing was dismissed on the grounds of overestimating wind speed over that region (not shown; cf. Figure 4b in Assmann et al. [2003]); however, the ECMWF winds, on the contrary, may be too slow over the same area in summer.

Sea ice speed $u$ is defined by a linear relationship as a function of the geostrophic wind $G$ and the mean ocean current $\bar{c}$ as $u = AG + \bar{c} + \varepsilon$ by Thorndike and Colony [1982], where $A$ is a complex coefficient that determines the scaling factor, $\varepsilon$ is the sum of other factors, and $u$, $G$, $\bar{c}$, and $\varepsilon$ are evaluated as complex numbers. The scaling factor is estimated to be $2-3 \times 10^{-2}$ from observations [Martinson and Wamser, 1990; Kwok, 2005]. Therefore, sea ice-ocean stress may become the dominant term in the sea ice momentum balance (see Equation (6)) when winds are weak. The calculation of the atmosphere-sea ice momentum exchange coefficient $c_u$ in Equation (31) is a parameterization that accounts for the stability of the atmospheric boundary layer [Hunke and Libscomb, 2004]; however, the sea ice-ocean drag parameter $c_{dw}$ in Equation (35) is a constant and is not well constrained [e.g., Martinson and Wamser, 1990; Perrie and Hu, 1997; Weatherly et al., 1998]. Kim et al. [2006] argued that CICE is sensitive to the sea ice-ocean drag parameter. In addition to the sea ice-ocean drag parameter, Timmermann et al. [2002], using an $s$ coordinate ocean model, reported sensitivity of sea ice drift to the turning angle $\theta$ between the sea ice and the wind field, whereas Reddy et al. [2007] found no such sensitivity with an isopycnal coordinate ocean model. For ocean models that adequately resolve the Ekman spiral, Hunke and Libscomb [2004] suggest setting $\theta$ to zero. In the current implementation, horizontal velocity grid points at the top ocean level are within the Ekman layer, and therefore, the default value $\theta = 0$ is used.

In general for this study, overall sea ice drift and areal distribution of summer coastal polynyas are well reproduced. Better representation of the eastern boundary conditions and wind forcing on the northern part of the domain would improve the simulated sea ice concentration over the deeper regions of the domain. Shelf and coastal regions, on the other hand, are prone to greater discrepancy due to being highly dynamical (Table 5). For the western Ross Sea in particular, this study shows that altering wind forcing in a subset of the region affects the simulated sea
ice concentration downstream (cf. Tables 5 and 7). Better resolved wind forcing and constraining the relative contributions of wind and ocean drag on sea ice would improve modelled concentration field.

IV. 1.2 Sea Ice Thickness and Mass Balance

Sea ice thickness, in terms of both domain averages and horizontal distribution, is well reproduced by the reference simulation. Modelled thickness field, produced thermodynamically at reasonable rates (Figure 22a, c), shows favorable reaction to dynamics (Figures 20, 22b, d). In winter, contrary to the simulated sea ice concentration, the thickness field better reflects the model response to atmospheric forcing that causes divergent sea ice cover (Figures 26-29(e, f)). Coastal regions, over which low sea ice concentrations are observed, consistently yield simulated thin ice. Effects attributed to wind forcing and sea ice flux at the open boundaries similarly influence the modelled thickness. Sea ice accumulation along the coast south of Cape Adare with thicknesses greater than observations; however, is not immediately explained by forcing and is possibly caused by inefficient export of sea ice out of the region, and consequently out of the domain.

Well-reproduced simulated sea ice thickness field implies confidence for calculations of sea ice mass balance. The sea ice volume budget, closed among the dynamic and thermodynamic sea ice states (Figure 23b), demonstrates the model’s ability to conserve mass and confirms the Ross Sea as an area of net sea ice production and export [Jacobs et al., 2002; Assmann et al., 2003; Kwok, 2005]. Freshwater equivalent of the mean sea ice export out of the domain is about 41 mSv. Assmann et al. [2003] calculated a ten-year (1988-97) mean modelled sea ice export off the continental shelf (described as the region within the 1000 m isobath) to be equivalent to 31.7 mSv. For this study, if all the net sea ice production over the continental shelf were exported, the flux would be 28 mSv. The value reported by Assmann et al. [2003] may be affected by overestimated sea ice thickness; however, the two fluxes are of the same order.

Horizontal distribution of sea ice growth and export rates determines the locations and magnitudes of important water mass transformations in the region. Polynya areas in southwestern Ross Sea, noted as regions of enhanced sea ice formation and export [Jacobs et al., 2002], contribute up to more than twice their area fraction to overall continental shelf sea ice production (Table 6, Figures 22a, c, 36b). Over the continental slope and abyssal Ross Sea, however, regional contribution to cumulative sea ice production is less than percent area occupied by these regions (Table 6).
Simulated sea ice production rates (Figure 36a), in agreement with previous regional modeling studies [Fichefet and Goosse, 1999; Assmann, 2003; Reddy et al., 2007], show that the Ross Sea continental shelf is an area of net growth and export.

On the Ross Sea continental shelf, simulated annual mean sea ice production over two years is 2.2 m over an area of 454,725 km$^2$. Using observed sea ice concentration along a linear exit gate extending from Cape Adare to Cape Colbeck, specifying a typical sea ice thickness of 0.5 m as reported by Jeffries and Adolphs [1997], and utilizing winds from NCEP-NCAR and ECMWF reanalysis products, Jacobs et al. [2002] and Kwok [2005] estimated mean ice production as 1.1±0.5 m and 2.5±0.8 m over 9 months within an area of 490,000 and 400,000 km$^2$ of the Ross Sea shelf, respectively. Assmann [2003] reported a 24-year (1978-2001) modelled mean annual growth of 1.5±0.24 m south of 70°S from Cape Adare to 150°W. Of this production, 2.57±0.37 m was west of 180° and 1.03±0.24 m to the east, with high rates taking place along the Ross Ice Shelf edge. For this study, mean annual sea ice production west and east of 180° south of the shelf break (Figure 14a) is about the same over areas of 207,275 km$^2$ and 247,450 km$^2$, respectively. Ross Sea polynya area sea ice production is divided roughly in half along 180° (Figure 22a, c). In the model domain, the shelf break is south of 70° and inclusion of low sea ice production from the lower latitudes (Figure 36a) would reduce the mean east of 180°.

Advection of sea ice northward off the continental shelf moderates the thickness distribution (Figure 20) and results in local melting that is a small fraction of the thermodynamically produced sea ice over the shelf (Figure 36b). Fichefet and Goosse [1999] reported 40% simulated local melting of mean 0.9 m thick sea ice on the southwestern corner of the Ross Sea. On a meridional transect along the Ross Sea polynya, Smith and Gordon [1997] found no significant sea surface dilution from meltwater late spring in mid-November. Simulated local melting over the continental shelf for this study is about 15%. The results from the z-t model described in Section II.3.2 show that the shelf ocean cannot sustain a purely thermodynamic sea ice cycle (Figure 35) due to the lack of local vertical water column heat balance as seen to occur in West Antarctic Peninsula (WAP) area [Smith et al., 1990].

IV.1.3 Ross Sea Fresh Water Budget

The annual fresh water budget for the Ross Sea (Figure 70), calculated from the simulated fields of ice shelf melt, sea ice formation and export, and atmospheric forcing, shows that atmospheric and ice shelf inputs of fresh water onto the Ross Sea continental shelf are 14.53 and 13.53 mSv, while fresh water of atmospheric, ice
Figure 70. Annual simulated fresh water budget for the Ross Sea. Fluxes (in mSv) are expressed as the fresh water equivalent of net ice shelf melt rate, net sea ice production rate (frazil and congelation ice only), net ice and snow export rate, and precipitation minus evaporation (P-E) from 2000 (a) and 2001 (b) from the reference simulation. Evaporation term includes condensation at the sea ice/snow surface and net sea ice exported includes snow-ice and ice formed by condensation.

Net ice shelf, and oceanic origin is extracted at 33.17 and 35.64 mSv for 2000 and 2001, respectively. Net ice production rates for the same two years, on the other hand, are 20.60 and 21.04 mSv, respectively. In the budget calculation, net sea ice production rate is the rate at which fresh water is extracted from the ocean, excluding snow-ice formation and vapor condensation at the sea ice/snow surface. The export term does not resolve sea ice types, and hence, is greater due to inclusion of sea ice incorporated into the pack directly from the atmosphere. The net fresh water export from the Ross Sea continental shelf, calculated as (net sea ice/snow export rate – precipitation minus evaporation (P-E) – ice shelf melt rate), is 18.64 and 22.11 for 2000 and 2001, respectively.

For the “inner” Weddell Sea, Timmermann et al. [2001] reported a 9-year average of 5.3 mSv freshwater extraction, calculated from simulated fields from a sea ice-ocean coupled modeling study. The part of their model domain over which the budgets were integrated is not confined to the continental shelf and includes open ocean where sea ice production and export processes are not as dynamic as the shelf areas. Net sea ice export and total fresh water input from P-E and ice shelf melt reported from the same study are 33.7 and 28 mSv, respectively. Although a direct comparison of numerical values from the Timmermann et al. [2001] study and the values reported here is not possible, it could be stated that the Ross Sea continental shelf is comparable to
Weddell Sea in that both shelves experience fresh water deficit due to sea ice processes and the rates are in the same order of magnitude.

IV.2 POLYNYA PROCESSES

Relatively low resolution and “land contamination” effect on the retrieval algorithm near the coast limit the use of satellite passive microwave data for polynya identification and size detection [Cavalieri et al., 1996; Dokken et al., 2002] and higher resolution retrieval algorithms for coastal regions have been proposed [Markus and Burns, 1995]. Contrast between brightness temperatures of land and ocean results in spurious sea ice concentration along the coastline, known as the “land-to-ocean spillover” [Cavalieri et al., 1997]. Along the expanding edge of the summer Ross Sea polynya, on the other hand, larger discrepancies between the SSM/I and AVHRR data have been reported [Zibordi et al., 1995]. Studies have used various criteria, such as sea ice concentration less than 75% [Massom et al., 1998], freezing rate greater than 1 m per month and sea ice concentration less than 70% [Marsland et al., 2004], and mean sea ice thickness less than 0.3 m [Smedsrud et al., 2006], as definitions alternative to that by World Meteorological Organization [1970], possibly due to these limitations. The WMO definition of a polynya “allows” up to 0.3 m sea ice cover, i.e. young ice, with no specific upper limit on the concentration [World Meteorological Organization, 1970]. In this study, 50% sea ice cover is found to be the lowest concentration value in the SSM/I winter data that yields a reasonable area for comparison (Figures 39, 43, 47) over the selected polynya regions (see Figure 14a).

IV.2.1 Polynya Size

The Ross Sea winter polynya signal in the SSM/I data appears as episodic, short-lived events occurring 8 to 11 times from April through September with areas in the 3-25x10^3 km^2 range (Figure 39b). Modelled polynya events match the timing of the observed openings well; however, the polynya areas are smaller than observed. Consistently during the simulated polynya events, the mean sea ice thicknesses are low, about 0.3 m or less (Figure 39a), wind magnitude is high (Figures 39c, 71), air temperature forcing shows warmer peaks (Figure 39d), there is net sea ice export (Figure 40c), congelation ice formation rate drops, frazil ice formation rate increases, and no local melting takes place (Figure 40b).

For the Ross Sea polynya area, winter wind and air temperature forcing are highly correlated to the observed sea ice concentration (Figure 71). Temporal match between
modelled and observed polynya events, therefore, shows model’s ability to respond to high-frequency forcing. The discrepancy in the winter polynya size, however, may be related to the magnitude of forcing and modelled sea ice growth and export as well as accuracy of the validation data. Fluctuations in the wind and air temperature forcing coincide and collectively drive the polynya. Plausible causal relationship is that the winds initiate the polynya and the ocean warms the air due to loss of insulation. Since there is no melting, winds determine the polynya size depending on the sea ice mass being dragged at the start of and during the event. The ocean model does not warm the air explicitly, and therefore, accurate growth rates during the polynya event are determined mainly by air temperature.

Terra Nova Bay polynya events in both simulations show similar progress and sea ice state features as those for the Ross Sea (Figures 72, 73). Forcing the Terra Nova
Bay area with ECMWF winds yield a very close match to the observed low sea ice concentration. For this case, simulated sea ice thickness, and growth and export rates are similar to those for the Ross Sea polynya area. Thus, inclusion of wind surges are not essential to create the observed open water fraction with the current sea ice model formulation. This is in agreement with the observation that intermittent coastal polynya formation is linked to synoptic winds [Zwally et al., 1985]. However, open water area alone does not create the latent heat polynya effect under moderate wind forcing. Surface sensible and latent heat fluxes are affected by the wind speed as formulated in the ocean model [Fairall et al., 1996]. Hence, higher wind speeds enhance sea ice formation by increased oceanic heat loss while exporting the newly formed ice more efficiently [Pease, 1987]. Due to export, sea ice is thinner and the rates of congelation ice formation increase.

Figure 72. As in Figure 71, but for the Terra Nova polynya.
IV.2.2 Growth and Export Rates

Simulated sea ice formation rates in the Ross Sea polynya area (Figure 40b) are consistent with the finding that congelation ice formation dominates the development of Ross Sea shelf pack ice [Jeffries and Adolphs, 1997]. In both years of the simulation, congelation, frazil, and snow-ice contribute 88.6, 11.1, and 0.3%, respectively, to total simulated sea ice volume within 200 km from the ice shelf edge (see Figure 14a). Sea ice core measurements within 800 km from the coast in the Ross Sea polynya area in May and June show the respective contributions from frazil and congelation ice types as 16.3 and 64.9%; however, no separate statistics are reported for the cores obtained close to the ice shelf edge [Jeffries and Adolphs, 1997]. Simulated snow-ice contribution is in agreement with the reported ice core measurements near the western border of the region carried out in late December and early January [Jeffries and Weeks, 1992]. Net growth (Figure 40b) and export (Figure 40c) rates do not exceed 3 and 4 cm d$^{-1}$, respectively, and maintain the modelled polynya with a thin ice cover and a thin layer of overlying snow as observed in austral winter [Jeffries
and Adolphs, 1997] and late austral autumn [Smith and Gordon, 1997]. Assmann [2003] reported 24-year mean modelled growth rates along the Ross Ice Shelf edge at 180° between about 2 and 5 cm d⁻¹ April through October, with a corresponding September mean sea ice thickness of 1.4-1.6 m. A modelled mean value of simulated sea ice export at 2.5 cm d⁻¹ has been reported by Fichefet and Goosse [1999].

Cumulative simulated sea ice production in the Ross Sea polynya area is about 4 m annually (Figure 36), which amounts to 0.5 m if spread over the entire Ross Sea continental shelf. An estimate of Ross Sea polynya sea ice production from Electrically Scanning Microwave Radiometer (ESMR) brightness temperature data over a 145,800 km² region extending 540 km along the Ross Ice Shelf edge is an equivalent of 1.6-2.8 m covering the continental shelf [Zwally et al., 1985]. Considering that these values are drawn from an area of high rates of sea ice production (Figure 22a, c), the eastern extent of which is not included in the selected RSP region, simulated production, which is affected by overestimated sea ice compactness north of the Ross Ice Shelf edge, is close to the lower bound of estimated range.

Simulated frazil ice formation for the Terra Nova Bay polynya contributes 27.1% of the two-year total sea ice volume, decreasing from 28.1% in the first year to 26.1% in the second (Figure 44b). Congelation ice makes up 68.0 and 71.3% of the simulated sea ice volume in the first and second years. Mean snow-ice formation is 3.3% which is in agreement with ice core measurements northwest of Terra Nova Bay that showed no significant snow-ice formation while the thickness of the snow layer varied between 1 and 23 cm [Jeffries and Weeks, 1992]. Mean sea ice thickness at these locations were reported to be between about 0.5 and 1.5 m with the ice cores comprising more than 95% congelation ice, showing the effects of downstream accumulation of sea ice swept east-northeast away from the bay area. Sea ice thickness measurements reported for winter (June 1995, May-June 1998) and summer (January 1999 and 2000) in the Terra Nova Bay are in the 0.1-0.3 and 0.3-0.7 m range, respectively (Figure 3) indicating a reasonable match between the model results and observations.

Simulated sea ice production in the Terra Nova Bay attains maximum cumulative rates of about 27 m yr⁻¹ near the coast and the formation rates decrease zonally towards the ocean (Figure 22a, c), averaging about 9 m annually over a 6825 km² region for the reference simulation and less than half of that when the polynya is forced by weaker ECMWF winds (Figure 45). Numerical estimates of cumulative sea ice production in the Terra Nova Bay polynya, using open water area from observed averages [Bromwich and Kurtz, 1984; Kurtz and Bromwich, 1985] or from a one-dimensional model of polynya width [Van Woert, 1999a; Fusco et al., 2002], range between 37.9
and 81.7 m annually. For the latter numerical method, frazil ice collection depth and sensible heat exchange coefficient determine the polynya extent to first order [Van Woert, 1999b]. Modelled rates reported from short-term integrations of a coupled atmosphere-polynya width model can be as high as 50 cm d⁻¹ [Gallée, 1997]. Ocean surface net heat loss used for these calculations are usually obtained by using data and bulk flux parameterizations and vary between 713 and 1517 W m⁻², up to an order of magnitude greater than estimated averages for the entire Ross Sea continental shelf [Budillon et al., 2000]. Negative heat flux thus obtained is applied over a contiguous area of ice-free ocean ranging from 1000 to 1300 km². However, neither model grid points nor SSM/I pixels in the Terra Nova Bay indicate cells that are 40% or less ice-covered during winter (cf. Figures 43b and 74b. Note also the decrease in observed Ross Sea polynya area with the >60% open water area criterion (cf. Figures 39b and 74a)). In the model implementation, any open water area in a grid cell that contains sea ice is treated as subgrid-scale by weighting the atmospheric heat exchange with sea ice concentration. Such a scaling distributes a fraction of the heat lost to the atmosphere over the whole grid cell, thus dampening sea ice formation rates over the ice-free area. Overall, simulated Terra Nova Bay cumulative production is equivalent to about 0.14 m of ice covering the Ross Sea continental shelf, comparing favorably to 0.2 m estimated by Kurtz and Bromwich [1985].

Another factor that may potentially decrease frazil ice production is the model formulation allowing for supercooling during the ocean model time step before the water temperature is restored to the in situ freezing temperature (Figure 8). Surface heat and buoyancy flux calculation, followed by potential temperature and salinity advection and diffusion are time-stepped before excess cooling is converted into a latent heat flux that forms frazil ice in the water column. Field measurements of the Arctic sea ice cover indicated no supercooling [Smedsrud and Skogseth, 2006] and a maximum of 0.02°C was observed in laboratory experiments [Smedsrud, 2001, cited in Smedsrud et al. [2006]]. In order to avoid any supercooling or allow a maximum amount, the model potential temperature field has to be checked continuously at the expense of increasing the computational cost to remove excess cooling in the form of frazil ice. Smedsrud et al. [2006] further argue for the importance of inclusion in sea ice model of a separate frazil ice class to account for the details of transition from open water to grease ice to consolidated ice pack, and also to accurately calculate the surface energy balance which is arguably a complex function of the classes and amount of sea ice present in the surface ocean layer [Kurtz and Bromwich, 1985; Gallée, 1997].
Figure 74. Simulated (black) and observed (red) total area of less than 40% sea ice cover. (a) RSP region and (b) TNB region from the reference simulation.

It should also be noted that the current implementation does not account for mass conversion between water and ice by treating formation and melting analogous to evaporation and precipitation, respectively. Cumulative production in the order of tens of meters annually is about two orders of magnitude more than maximum evaporation observed for the world ocean [Pickard et al., 2007] and would result in sea surface height deficits that may need to be accounted for over wind-driven polynya areas during periods of enhanced sea ice production.

Tuning or improving parameterizations that would potentially increase frazil ice formation rates in polynya areas should be accompanied by a better constraining of the collection thickness which is proportional to polynya width and opening time [Pease, 1987]. For the simulated polynya events, winter sea ice concentration in excess of observations may moderate the formation rates by letting the ocean keep heat that would otherwise be lost to the atmosphere in the form of sensible and latent heat, i.e. dominant terms in the winter surface heat budget [Smith and Klinck, 2002]. However, since simulated sea ice thickness is well reproduced in the polynya
areas, conductive heat flux through thin sea ice, treated as an uncertainty in sea ice production estimates from surface energy balance [Kurtz and Bromwich, 1985; Van Woert, 1999b], compensates for part of the error term in the ocean surface heat flux.

IV.2.3 Ocean Response to Polynya Dynamics

During the winter months, there is no significant simulated basal melting in the Ross Sea and Terra Nova Bay polynya areas (Figures 39b and 43b) indicating that sensible heat, if any, made available by convective, advective, or diffusive processes to the ocean surface layer and subsequently transferred to the atmosphere can only be involved in slowing down or preventing sea ice formation in the polynya area. Other than persistent warm MCDW intrusions onto the continental shelf, the most probable source of oceanic heat to polynya areas during winter [Pillsbury and Jacobs, 1985; Jacobs and Comiso, 1989; Gordon et al., 2000; Dinniman et al., 2003], data from moored temperature sensors near the Ross Ice Shelf led Jacobs and Comiso [1989] to argue that water remaining below winter mixed layer during the freezing season, at a degree or more above surface freezing point and within the reach of the winter vertical circulation, can provide heat to the surface even if warm water intrusions onto the shelf do not persist, and as such contribute to maintaining low ice concentrations. In both polynya regions, the winter convective layer in the reference simulation reaches deep in the water column below the surface where majority of thermodynamic ice production takes place (Figure 22a) and warmer water masses exist at intermediate depths in the vicinity of the polynya area (Figures 41a, c and 50a, c). Oceanic heat transferred laterally into the polynya area from the surrounding warmer water is about the same magnitude per area for both polynya regions (Figures 40e and 44e). The main difference between the two regions, however, is the relative contribution of lateral heat influx to water column heat budget.

Oceanic heat gained laterally from out of the region in the Ross Sea polynya area compensates for the bulk of the heat loss to the atmosphere and moderates frazil ice formation; whereas, for the Terra Nova Bay polynya area, heat gained from the temperature restoration of supercooled water comprises about half of the positive heat flux. Although low sea ice concentrations in both polynya regions are driven by divergent motion forced by the winds, such distinction in ocean response causes the Ross Sea winter polynya to conform more to the "sensible heat" definition. Negative ocean surface heat balance in the Terra Nova Bay area results in enhanced supercooling and frazil ice formation due to insufficient lateral heat gain and high rates of ice formation in turn causes the surface convective layer to reach the sea floor. The relationship
between sea ice formation rates and oceanic heat influx is further demonstrated with
the ECMWF simulation where the winter mixed layer remains shallower (Figure 50b, d) as a result of low formation rates due to weaker winds. Terra Nova Bay winter polynya for this case is even more sensible heat-maintained (Figure 48e) than the Ross Sea counterpart.

IV.3 WATER MASS DISTRIBUTION AND TRANSFORMATIONS

IV.3.1 Modelled Hydrography and Circulation

Simulated Θ-S distribution of the Ross Sea capture the main observed hydrographic characteristics of the region. AASW, which reflects a distinguished seasonal signal, is more abundant over the continental slope and, over the continental shelf, its volume decreases substantially west of 180° (Figures 32 and 32) in agreement with observations recently compiled into a comprehensive temperature-salinity database [Stover, 2006]. The salinity distributions at 400 and 500 m (Figures 53 and 55) show that modelled shelf waters are slightly fresher along a meridional channel between 170°E and 180° at both depths and slightly saltier at 400 m along and north of the Ross Ice Shelf edge west of 175°E, while otherwise indicating a very close match with the observed distribution [Stover, 2006]. Temperature range of the water in the ice shelf cavity (Figures 32a and 32a) are within the observed range [Jacobs and Giulivi, 1998; Gouretski, 1999] and indicate the presence of AASW entering the cavity and cooling after thermal exchange with the ice shelf base. HSSW, confined to and spread over most of the shelf region west of 170°W; LSSW, found in patches over western Ross Sea shelf and along the Ross Ice Shelf edge between about 165°E and 165°W with a northern extent well south of the shelf break; and MSW/AABW appearing over the entire shelf area [Stover, 2006] are well reproduced by the reference simulation (Figure 66). Zonal (Figure 52) and meridional (Figure 42) model salinity cross sections indicate the influence of sea ice formation rates on the density structure [Zwally et al., 1985; Buffoni et al., 2002] by driving the regional and seasonal variability, respectively. Features important for the heat budget of the Ross Sea shelf, namely the Antarctic Slope Front [Jacobs, 1991] and related warm CDW inflows [Jacobs and Comiso, 1989] are also well simulated (Figures 41 and 57).

Model circulation (Figures 21c, f and 24) agrees with observed characteristics in that it is mainly barotropic, constrained to follow topography, as suggested by current meter measurements [Picco et al., 1999]. Westward coastal flow splits west of 160°W with branches southwestward following the Ross Ice Shelf edge [Pillsbury and
Jacobs, 1985] and northwestward along the continental shelf break [Keys et al., 1990; Picco et al., 1999]. Shelf circulation includes two cells of anticyclonic transport over the shelf in agreement with patterns deduced from observations [Pillsbury and Jacobs, 1985; Locarnini, 1994], with the western cell also discerned from current meter measurements [Jaeger et al., 1996; Picco et al., 1999]. West of the western limb of the anticyclonic cell, transport is northward along the western coastal boundary off Victoria Land [Picco et al., 1999]. Flow enters into the ice shelf cavity in southwestern [Lewis and Perkin, 1985; Stover, 2006] and eastern Ross Sea around 175°W [Pillsbury and Jacobs, 1985] and outflows are located in the central [Jacobs et al., 1970; Pillsbury and Jacobs, 1985; Picco et al., 1999; Stover, 2006] and southeastern Ross Sea east of Roosevelt Island [Jacobs et al., 1970; Hellmer and Jacobs, 1995].

Simulation results consistent with observational studies of Ross Sea shelf circulation along with simulated Θ-S distribution in general agreement with available data compilations provide a material basis for quantitative estimates of water mass transformations and bottom water formation in the study area.

IV.3.2 Ross Sea Shelf Waters

Modelled LSSW net transport is well developed along central and eastern Ross Sea east of 180° (Figure 66c, f), where about 80% of its climatological volume is concentrated [Stover, 2006], although the eastern extent of the simulated volume is east of that for the climatology. West of 180°, where LSSW is observed to be relatively less abundant [Whitworth et al., 1998; Stover, 2006], net transport of this water mass is minimal and scattered. Net onshelf LSSW volume flux northward across the Ross Ice Shelf edge, which also includes Ice Shelf Water (ISW) by definition for the current analysis (Figure 15), is a result of HSSW interacting with the ice shelf base after it flows into the cavity in western Ross Sea (Figures 66b, e and 68a) as argued from observations [MacAyeal, 1984; Smethie and Jacobs, 2005].

The salinity, distribution, and circulation of HSSW on the western Ross Sea shelf (Figures 51, 52, 55, 56) are influenced by brine rejection from sea ice formation, especially in coastal polynya areas [Zwally et al., 1985; Jacobs and Giulivi, 1998; Van Woert, 1999a; Budillon and Spezie, 2000]. Studies on quantification of HSSW production in western Ross Sea focus on the Terra Nova Bay polynya area due to enhanced sea ice production discerned from increased heat loss over the ice-free ocean surface. HSSW production from the reference simulation, calculated as the difference between yearly mean inflow (0.88 Sv) and outflow (1.02 Sv) of HSSW for the selected region, is 0.14 Sv (Table 8). Simulated HSSW production takes place mainly in winter
with maxima greater than 0.5 Sv, while near zero during the summer months (Figure 68c) as suggested by observations in December and February [Budillon and Spezie, 2000].

About 1 Sv of HSSW production and/or transport in the Terra Nova Bay polynya area is a common result reported from studies that use various methods for estimation. Surface heat balance-one dimensional wind-driven polynya models use salt production from sea ice formation to transform LSSW to HSSW, assuming a sustained conversion between salinities 34.5 and 34.8 [Van Woert, 1999a; Fusco et al., 2002]. Manzella et al. [1999] reports about 0.85 Sv of transport only from March through November based on current, temperature, and salinity observations. Buffoni et al. [2002], on the other hand, used a two-dimensional model, with specified sea ice production (45 and 90 m yr⁻¹) and horizontal velocity (0.2-0.4 m s⁻¹), to estimate conversion from incoming source water (S between 34.65 and 34.70) to HSSW (S > 34.77) and estimated a net northeastward transport ranging from 0.34 Sv to 1.23 Sv. On the other hand, an analytical calculation for the Arctic, for which the dense shelf water production from all coastal polynyas is estimated about 0.7-1.2 Sv [Cavalieri and Martin, 1994], yielded a value of 0.04 Sv dense water formation per 100 km of coastal polynya [Chapman, 1999].

Realistic estimation of HSSW formation and net transport in the Terra Nova Bay polynya area depends on correctly identifying the ambient water masses that are salinized by the brine released during sea ice formation. Observations in austral summer in the area indicate that the water column is occupied by HSSW below the warmer and fresher surface layer which deepens from 50-100 m in December to 150 m in February [Budillon and Spezie, 2000]. During the period of observations, there was no clear presence of MCDW in the area and although the summer polynya was intermittently active during katabatic events, HSSW production was not hypothesized due to low rates of estimated salt flux and lack of the surface water preconditioning phase. The thickness of the low salinity (34.5 ≤ S < 34.7) MSW/AABW layer in the Terra Nova Bay polynya area from a Θ-S climatology of the Ross Sea is between 25 and 75 m while there are no significant layers of LSSW and high salinity (S ≥ 34.7) MSW/AABW present [Stover, 2006]. The thickness of the high salinity MSW/AABW layer to the south of the Drygalski Ice Tongue in the same data product is between 75 and 100 m, which may provide additional fresher source water for HSSW production upstream. The model results show that HSSW and MSW/AABW constitute the majority of transport through the Terra Nova Bay water column which provides a sink for small amounts from each of MCDW/LCDW, MSW/AABW, LSSW (Table 8). Considering
that during winter the surface convective layer reaches the sea floor after AASW is cooled and salinized (Figure 50a, c), simulated “new” HSSW formation rates not in the order of 1 Sv, as frequently reported in literature, is consistent with the absence of such volume of source water masses in the polynya area.

IV.3.3 The Role of Terra Nova Bay

The net effect of simulated Terra Nova Bay polynya is salinization of HSSW as argued by Jacobs et al. [1985]. Kurtz and Bromwich [1985] estimated that up to 50% of salinization of HSSW could be due to sea ice formation in the area. The extent of the influence of Terra Nova Bay sea ice processes reaches far over the larger westernmost shelf area, including the region north of Ross Island and the Drygalski Trough, where differences in salinity are up to 0.2 (cf. Figures 51a, c, 52, and 53-56) when forced by different winds. Temperature of the fresher water column resulting from weaker winds is higher (Figure 51b, d), placing the water mass in the MSW/AABW category. While northward transport rate of HSSW and mid-depth to bottom salinity over a considerable area of the western shelf decrease in response to weaker Terra Nova wind Bay forcing, south- and eastward net transport of HSSW north of McMurdo Sound and east of Ross Island, respectively, increases (cf. Figures 66b, e and 67b, e) and the water mass along the western part of the ice shelf edge is saltier at 400 m (cf. Figures 53 and 54). Hence, HSSW volume is not maintained along the coast and direction of transport reverses south of Drygalski Ice Tongue. According to temperature, salinity, and current measurement studies [Pillsbury and Jacobs, 1985; Picco et al., 1999], the zonal salinity gradient in the Ross Sea continental shelf drives the thermohaline circulation [Jacobs and Giulivi, 1998]. High rates of sea ice formation, as in the reference simulation, maintain the volume and northward coastal transport of HSSW while with weaker winds the HSSW layer erodes, MSW/AABW constitutes the majority of volume flux through the bay area, MCDW/LCDW intrusions take place (Table 9) unlike what is observed [Budillon and Spezie, 2000], and as a result of weakening zonal density gradient, the southward limb of the western anticyclonic gyre [Locarnini, 1994] slows down (Figure 67a, c). Increased stratification over the western part of the shelf may also disrupt possible dense subsurface water upwelling that was suggested by Killworth [1974] as a mechanism for maintenance of the zonal salinity gradient; however, only a north-south variation in brine influx due to differences in sea ice formation was assumed in that study.
IV.3.4 Modified Shelf and Antarctic Bottom Waters

Formation rate of AABW, for which highly variable estimates have been reported from geochemical tracer and modeling studies, is not well constrained [Jacobs, 2004]. The percent contribution to AABW production from the Pacific sector of Antarctica, representative of the Ross Sea, is reported to be between 20 and 33% [Locarnini, 1994; Nakano and Sugino, 2002], where estimates of the Antarctic total are in the 8-41 Sv range [Jacobs et al., 1985; Orsi et al., 1999; Jacobs, 2004]. Such a range of variability stems from different definitions used for AABW and the variety of methods involved in the analyses. Considering that the transport magnitudes are only several Sv on a regional level, the results may be sensitive to discrepancies in water mass definitions. For this study, simulated deep water masses were classified by ranges of potential temperature and salinity (Figure 15) reasonably consistent with recent studies that utilized neutral density surfaces [Jackett and McDougall, 1997] to define Antarctic water masses involved in mixing processes over the continental margin [Whitworth et al., 1998], bottom water production [Orsi et al., 1999; Orsi et al., 2002], and to provide a comprehensive Θ-S data product for the Ross Sea [Stover, 2006].

Simulated MSW/AABW constitutes most of the dense off-shelf flow (Table 8). Small HSSW and LSSW transport across the shelf break takes place where the analyzed section is shallower than 1000 m (Figure 66). Two-year average MSW/AABW transport is 2.23 Sv for the reference simulation. Orsi et al. [2002] deduced 3.2 Sv AABW bottom layer abyssal input from the Indian-Pacific sector from CFC measurements. When top and middle layers of AABW are included in the total (8.1 Sv) and considering the 3:2 ratio for relative contributions of the Indian and Pacific sectors [Jacobs, 2004], Ross Sea AABW transport is about 3 Sv from that study. Although estimates for circumpolar AABW output cover a wide range, simulated values agree more favorably with the lower bound of the range.

The ECMWF simulation shows an average of 0.15 Sv decrease in AABW outflow over two years and a decreasing trend from the first year to the next (Table 9). This is consistent with the dilution and weakening circulation in the western shelf regime for the same simulation. Although the duration of the simulation is not long enough for a conclusive statement, the effects of reduced sea ice production on the water mass properties, distribution, and transformation may be expected to continue. Toggweiler and Samuels [1995] argue that ocean general circulation models may tend to cause freshening of deep and bottom waters due to circulation deficiencies and that sea ice production may not have an important effect on the salinity of AABW. The ECMWF
simulation results indicate such a dilution and weakening circulation, however this is due to altered sea ice formation rates in the Terra Nova Bay. There also is a slight decrease in off-shelf transport of saltier AABW in the second year of the ECMWF case compared to the reference simulation (Figures 64b and 65b).

The reference simulation reproduces continental shelf and slope features that are important in AABW formation, namely, LSSW/ISW extending towards the shelf break (Figure 75), HSSW filling the depressions and extending north over the western continental shelf (Figure 51a, b), and the “V-shaped” slope front (Figures 51 and 75) where ambient waters throughout the water column mix and descend deep [Jacobs, 2004]. However, mechanisms also deemed important for processes related to AABW formation and export, namely tides [Beckmann and Pereira, 2003; Gordon et al., 2004] and dense downslope gravity plumes [Baines and Condie, 1998; Bergamasco et al., 2002; Gordon et al., 2004] are not modelled in the current implementation.

Figure 75. Simulated potential temperature (a) and salinity (b) cross sections along the eastern edge of the PLY region (see Figure 14b), but extending south to the ice shelf edge and north past the shelf break (looking east). Snapshots are for February 11, 2001 from the reference simulation.
CHAPTER V
CONCLUSION

The Ross Sea model provided a framework to investigate the coupled glacial and sea ice-ocean system in a region of scientific interest due to features such as the presence of a large ice shelf cavity, recurring polynyas, wide and abruptly sloping continental shelf over which the densest shelf waters in Antarctica are formed, and as a result of their combined effects, an important source region for Antarctic Bottom Water. Integration of the model processes at a resolution high enough to resolve small scale polynyas and at an horizontal extent wide enough to cover the shelf and abyssal Ross Sea made it possible to analyze the basin wide effects of regional dynamics, as reported in particular for the Ross Sea and Terra Nova Bay polynyas.

In terms of simulating the large scale seasonal sea ice concentration field, the model performs better over the continental shelf and coastal areas, namely, away from the open boundaries in the vicinity of which discrepancies with the SSM/I data are more pronounced. Reproducing sea ice production and export rates over the continental shelf and polynya areas in good agreement with the estimates obtained from observations is essential to have reasonable confidence in calculating shelf water mass budgets that are sensitive to the location and amount of sea ice produced. Accurate estimation of water mass budgets and the transformation rates are important in terms of assessing their importance for the larger scale thermohaline circulation and how it may be affected by climatic changes in the local and global scale.

The simulations showed that warming of the atmosphere would decrease sea ice formation rates over polynya regions where frazil and congelation ice formation rates, which depend on the temperature difference at the air-sea interface, are high throughout the winter. It is also shown in the context of Terra Nova Bay that the amplification of ocean surface heat loss by the winds on a local scale have a larger area of influence. Therefore, weakening wind field on the continental shelf or a changes in the frequency and speed of the katabatic winds due to warming would affect the salinity of the bottom water formed in the Ross Sea.

The model successfully recreates the differences in sea ice dynamics between the east and west parts of the Ross Sea, which in turn maintains the simulated zonal salinity gradient that localizes HSSW to the western and LSSW to the eastern shelf areas, respectively. Although sea ice formation and melting rates over the abyssal part of the domain are well reproduced, distribution of the pack is influenced by the wind
pattern close to the open boundaries and sea ice transport across these boundaries. For the eastern open boundary, in the absence of accurate, high-frequency time series of sea ice thickness and velocity, including this highly dynamic region within the domain or imposing sea ice thickness from well-verified circumpolar sea ice model output would improve the simulated sea ice thickness field along the eastern part of the continental slope. The improvement would also reflect in the abyssal region spanning the southwest corner of the Ross Gyre. A similar potential improvement would also apply to the northern boundary, which is located to the south of the seasonal maximum sea ice extent, and as such, is sensitive to the direction of the winds driving northward sea ice export and to the sea surface temperature imposed at the open boundary for summer retreat.

Formation of dense, saline shelf water mass (HSSW) and its subsequent interaction with warmer MCDW and colder ISW/LSSW to produce MSW/AABW are driven by coastal polynya processes in the western Ross Sea. The model performance in simulating the Ross Sea polynya, which dominates the brine input in the region [Zwally et al., 1985], is seasonally determined. Winter sea ice concentrations are overestimated, possibly due to shortcomings inherent in synoptical wind forcing and in the sea ice model’s formulation of horizontal versus vertical distribution of mass. Considering that the sea ice volume, i.e. the grid cell average thickness, is well represented in the polynya regions, improvements in the treatment of formation and consolidation of frazil ice, which are ongoing processes throughout the winter in leads and polynyas as opposed to regions of more compact sea ice cover, may result in a better representation of the winter Ross Sea polynya area sea ice concentration. The spring and summer Ross Sea polynya, on the other hand, is simulated successfully in terms of timing and location. A reliable simulation of the sea ice formation rates makes the calculation of relative contributions from sensible and latent heat components to the formation and maintenance of the Ross Sea polynya possible. The winter polynya is wind-driven with the oceanic heat continuously moderating formation rates. This mechanism thus maintains a thin sea ice cover over the polynya area which is more efficiently exported during winter. As a result of the lack of thickening and accumulation, the polynya area sea ice cover is conditioned over the winter for rapid melting in late spring and summer when freezing is discontinued by atmospheric warming.

Especially during winter, the thermodynamics and dynamics of the sea ice pack are dominated by the atmosphere. In the absence of an atmospheric model component, it is essential for the forcing fields to adequately resolve relevant temporal and spatial scales of sea ice dynamics. In particular to polynya modeling, this includes episodic,
mesoscale wind events such as storms and katabatic surges. Substitution of observed wind forcing for a small subset of the domain, i.e. the Terra Nova Bay, substantially improves polynya behavior in terms of sea ice production and export, which in turn resulted in further-reaching, non-local influence on water mass distribution. It was shown that wind amplification on sea ice production was required to reproduce the effects of Terra Nova Bay polynya on the water column, i.e. maintaining, salinizing, and producing HSSW. The results imply that for regions that experience highly dynamic sea ice processes forced by the atmosphere, unrealistic wind forcing adversely affects the behavior of the simulated coupled sea ice-ocean system.

The overall performance of the coupled model over the Ross Sea continental margin in terms of seasonal sea ice dynamics, ocean circulation, and water mass distribution is in favorable agreement with the available observations which makes it possible to quantify water mass transformations with high spatial resolution and frequency that reflect the details of the seasonal signal. AABW formed and transported off-shelf as calculated from this study agrees with the quantities (and also supports the adequacy of the techniques used to obtain such quantities) reported by recent studies estimating contribution of the Ross Sea to AABW production.

The current modeling effort involving sea ice and ocean models that well represent the complexity of both systems at high spatial resolution comes at the expense of increased requirement for computational resources. Additional complexity involving tidal dynamics, which are argued to be important for the sea ice and ocean dynamics in polar regions [Padman and Kottmeier, 2000; Polyakov and Martin, 2000; Padman et al., 2003; Beckmann and Pereira, 2003; Koentopp et al., 2005], and coupled models of downslope dense gravity plumes [Harvey, 1996], one of the main mechanisms for transporting AABW off-shelf, may be required to improve model behavior and water mass budget estimates. Additionally, longer term simulations that are beyond the scope of this study would identify interannual and interdecadal variability in the system at temporal scales relevant to its sensitivity to changes in the global climate.
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Typeset using \LaTeX.